Volume 30 SEPTEMBER, 1960 Number 3

# Journal of Sedimentary Petrology

A Publication of the Society of Economic Paleontologists and Mineralogists
a Division of
The American Association of Petroleum Geologists

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*The term graywacke.* By F. J. Pettijohn.

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 353–369
FIGS. 1–12, SEPTEMBER, 1960

# PETROLOGY OF TAMWORTH GROUP, LOWER AND MIDDLE DEVONIAN, TAMWORTH-NUNDLE DISTRICT, NEW SOUTH WALES[1]

KEITH A. W. CROOK
University of New England, Armidale, N.S.W.[2]

## ABSTRACT

The Tamworth Group is dominantly of deep-water origin and consists largely of graywacke rudites, graywackes, and argillites. The latter usually contain radiolaria. Lithic labile graywackes are dominant and consist chiefly of andesitic detritus (rock fragments, plagioclase, ferromagnesians, chlorite, and accessories) with minor rhyolitic, granitic, and older sedimentary components. They are thought to have originated in complex island arcs.

Minor feldspathic labile graywacke and argillaceous limestone occurs. The former is thought to be largely of andesitic origin, being redistributed crystal tuff with admixed plutonic quartz. The limestones, small silty lenses, are probably the result of calcification during early diagenesis.

The feldspar in most of the group is now albitised, but relics of andesine-labradorite remain occasionally.

The Drik-Drik Formation low in the sequence is largely coarse keratophyric detritus, heavily hematitised and scarcely transported. It may be a shallow-water deposit derived from the underlying Copes Creek Keratophyre, or from contemporaneous volcanics on an adjoining land-mass.

## INTRODUCTION

The Tamworth Group, the lowest unit in the Devonian-to-Permian Tamworth Trough of northeastern New South Wales (fig. 1), reaches a maximum thickness in excess of 8000 feet. Its stratigraphy has been described by the author (Crook, 1960a) and is summarised in table 1. Sedimentary structures suggest that all units except the Drik-Drik Formation are deep-water turbidite deposits. The only previous petrological work on the Tamworth Group is that of Benson (1913, 1915a, 1915b, 1918).

The data presented are the result of hand specimen and thin section study; the material is housed in the museum of the Geology Department, University of New England. Specimen numbers refer to the University of New England Collections.

Refractive indices were determined by the immersion method, the R. I. of the oil being measured on an Abbe refractometer. Determinations on feldspars may be taken as correct to $\pm 0.002$ and those on pyroxene to $\pm 0.004$. Data are incorporated in table 2.

The number of slides examined during the study (54) is, in the author's opinion, suffi-

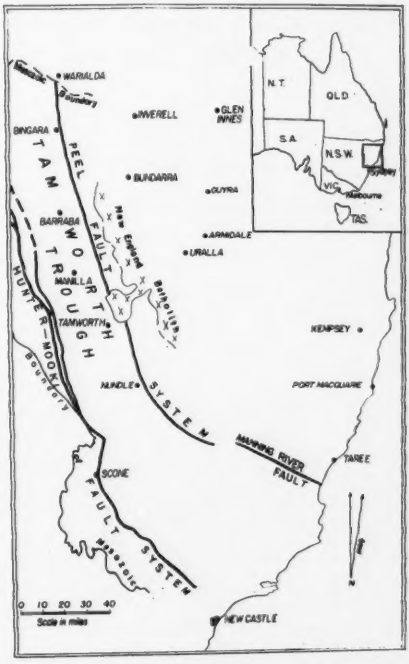


FIG. 1.—Location map (data from Voisey, 1959).

[1] Manuscript received October 29, 1959.
[2] Present address—Dept. of Geology, University of Alberta, Edmonton, Alta., Canada.

TABLE 1.—*Subdivisions of Tamworth Group*

| | | Eastern and Southern Region | Northern Region |
|---|---|---|---|
| DEVONIAN | MIDDLE | Yarrimie Formation (Dmy)<br>Silver Gully Formation (Dms) | Yarrimie Formation (including Levy Graywacke Member (Dmlg).)<br>Silver Gully Formation (Dms) |
| | LOWER | Wogarda Argillite (Dlw)<br>Drik-Drik Formation (Dld)<br>Cope's Creek Keratophyre (Dlc)<br>Pipeclay Creek Formation (Dlp)<br>Hawk's Nest Beds (Dh) (Stratigraphic position unknown) | Seven Mile Formation (Dls) |

cient only to give the general trend of the composition of the sediments. It is quite inadequate to give a clear picture of the variation in any formation along or across the strike. Fortunately the arenites throughout the sequence are, for the most part, remarkably uniform in composition, and this minimises the dangers of generalising on the data available.

Only three specimens have been modally analysed: two are lithic graywackes, the third is a feldspatholithic graywacke. They fall within the range of variation of the arenites of the overlying Parry Group and will be included in the discussion of these rocks to be published shortly. Visual estimation suggests that most Tamworth Group arenites would fall within the same range of variation.

In the descriptions following, diagenetic modifications are deliberately treated rather briefly. They will form the substance of a forthcoming communication on the diagenesis of the Tamworth and Parry Groups.

The penecontemporaneous igneous rocks included within the Tamworth Group (spilites, albite dolerites, keratophyres, and quartz keratophyres) will not be described. They have been very fully described by Benson (1915a, 1918).

### TERMINOLOGY

Terminology follows that of Pettijohn (1957) except for the classification of arenites where that of Packham (1954), as expanded by the author (Crook, 1960b), is used.

The following qualitative categories of sorting and roundness, based on visual estimation, have been adopted.

TABLE 2.—*Refractive index determinations* ($n_Y$)

| Stratigraphic Unit | Specimen Number | Pyroxene | Detrital Feldspar | Albitised Feldspar | Feldspar replacing Ca-feldspar |
|---|---|---|---|---|---|
| Dmy | R901 | | | 1.533 | |
| | R904 | | 1.541 | | |
| | R909 | | 1.553 | | 1.533 |
| | R920 | | 1.552 | | |
| | R922 | | 1.559 | | |
| | R923 | 1.695 | | | |
| Dmlg | R927 | 1.696 | | 1.534 | |
| Dms | R929 | | 1.553 | | 1.533 |
| | R932 | | | 1.534? | |
| Dms? | R954 | | 1.553 | | |
| Dld | R937 | | 1.558 | 1.534 | 1.533 |
| | R948 | | | | 1.534 |

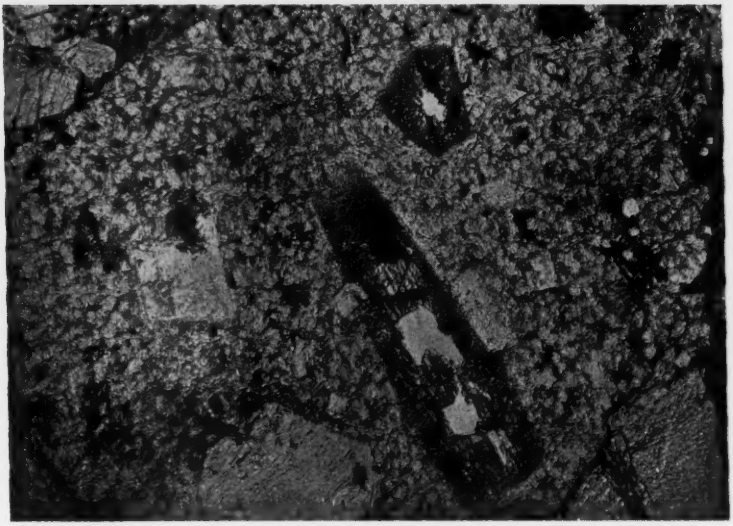

FIG. 2.—Albitised volcanic fragment showing "pseudo-clathrate" texture (×66). R828, a lithic labile graywacke from the Pyramid Hill Arenite, Member R, Parry Group.

very well sorted
well sorted
moderately sorted
poorly sorted
very poorly sorted

* * *

very angular
angular
subangular
subrounded
rounded
well rounded
very well rounded

Sphericity has been assessed visually using the chart of Krumbein and Sloss (1951, p. 81).

Most of the textural terms used to describe igneous rock fragments are taken from Johannsen (1939). The term *pseudo-clathrate* is used to describe a rock-type having a texture superficially similar to clathrate (which is characteristic of leucite rocks). The pseudoclathrate texture (fig. 2), like the clathrate, resembles a section through a sponge, but the minerals involved are fine chlorite moulded about ovate albite crystals. No genetic relationship with the leucite rocks is implied.

Two other types are termed *textural andesite* and *textural basalt*. The former is porphyritic in feldspar, rarely vesicular, and possesses a fine cryptofelsitic greenish groundmass (fig. 3). The textural basalt is holocrystalline and frequently subtrachytic, with a tendency to have intergranular chlorite and opaques (fig. 4).

## PETROLOGY

### Yarrimie Formation (Dmy)

This formation consists dominantly of hard, cherty black and white banded siliceous radiolarian argillites, which locally become sufficiently rich in silica to warrant the name "chert." They contain small lenses of argillaceous limestone, which may contain abundant radiolaria. Near the base of the unit the argillites are dark green and radiolarian; locally higher in the sequence bright green argillites occur. Approaching the top of the unit, mudstone interbeds appear and become more prominent upwards.

The formation contains many mappable arenite beds; one is designated as a separate

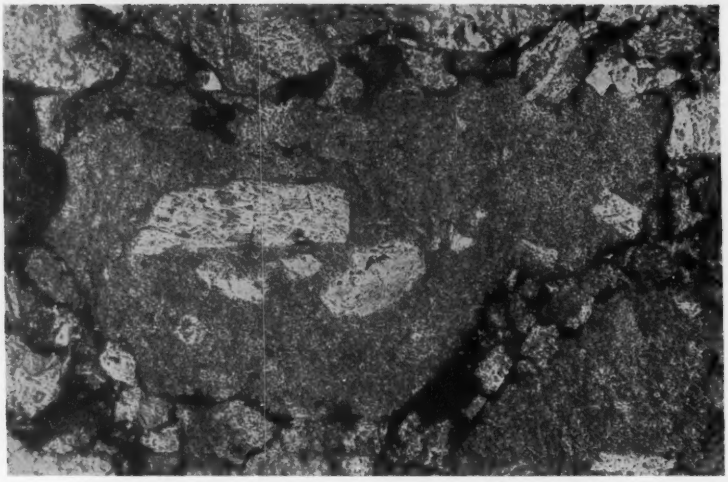

FIG. 3.—Volcanic fragment of the type termed "textural andesite" (×82.5). R824, a lithic labile sandstone from the Pyramid Hill Arenite, Member 6, Parry Group.

member. They are all lithic labile graywackes and are very poor in quartz (maximum 5 percent). Occasional feldspathic graywacke interbeds occur within the argillites. All arenites are well indurated.

*Lithic labile graywackes.*—(R901, R903-5, R907, R909, R915, R917-25) These arenites are usually of medium to coarse grain and

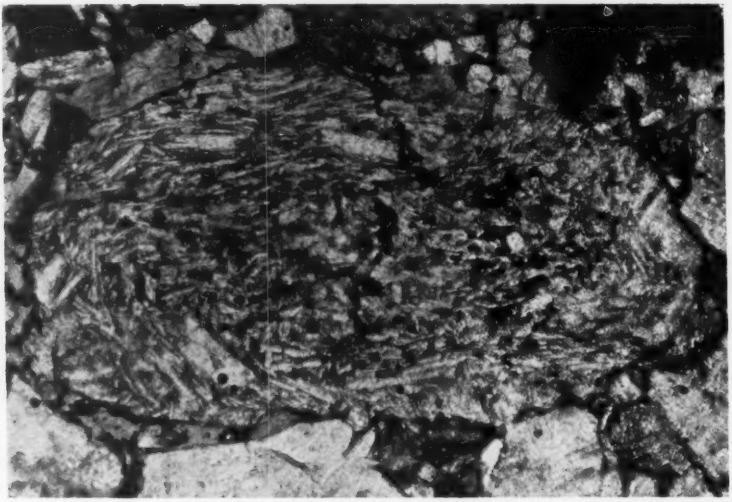

FIG. 4.—Volcanic fragment of the type termed "textural basalt" (×60). R867, a lithic labile graywacke from the Garoo Conglomerate Member, Parry Group.

may show graded bedding (fig. 5). One fine rudite is included. Sorting is generally moderate to poor, although some are somewhat better sorted. Roundness is very variable, mostly from subangular to rounded. Sphericity varies widely, but mostly in the range 0.5–0.8.

These rocks are usually dark gray or dark green and consist dominantly of volcanic rock fragments, apparently of andesitic affinities, several textural types usually being present. Plagioclase, usually albite, is prominent, and patches of chlorite occur sporadically. Minor amounts of quartz and accessories are usually present. Matrix is either subordinate or fairly prominent (fig. 6), and a cement is usually present. Detrital grains are usually in close contact, indicating a high degree of compaction.

*Quartz* grains are usually simple, often unstrained, but at times with wavy extinction. Rarely they are composite, some with sutured junctions between components (R904). Some have small amounts of orthoclase associated, indicative of granitic derivation (R901, R904, R907). Fluid inclusions are ubiquitous and frequently form trails. Obscure high relief and acicular inclusions are frequent. Minerals forming inclusions are brown tourmaline (R903), apatite (R907), chlorite (R918), and zircon. Obscure striae appear on a few grains (R922, R925). Embayed grains occur only in R904, but in several cases the grains appear to have suffered marginal corrosion in place. Broken grains of high spericity and roundness occur in R918. Quartz is replaced by pyrite in R922.

*Feldspar* is usually highly sodic but includes a little clouded orthoclase. Refractive indices (table 2) indicate a composition of $An_{2-3}$. This is not the original composition, as R904 and R907 contain more calcic plagioclase ($An_{18.5-53.5}$) veined with albite ($An_2$) or rarely by a zeolite (R904). R909 and R924 retain unaltered calcic plagioclase. The albite is usually slightly clouded and flecked with sericite. Occasional chlorite blebs are included. It may be replaced by epidote, carbonate, prehnite, and sericite (R915, R924) or by carbonate and sericite (R919). Albite twinning, though present, is rarely sharp and many grains are untwinned. The more calcic feldspar is clear, sharply twinned, and may be slightly sericitised.

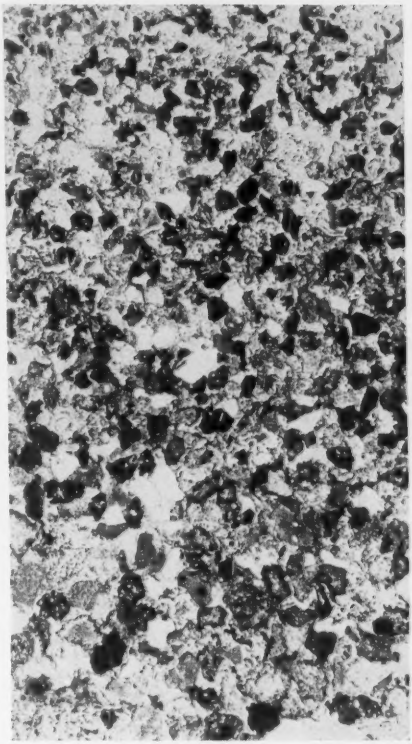

FIG. 5.—Lithic labile graywacke (R909) from Yarrimie Formation, showing good graded bedding (×7.5).

Clear untwinned feldspar, with 2V(+), large, occurs in R907, R920, and R924, in quantities insufficient for composition determination. Its R.I. appears to be close to that of quartz, and its close resemblance to this mineral is enhanced by its lack of alteration and occasional small fluid inclusions.

*Pyroxene* occurs in several specimens but is never abundant. It has 2V(+), large, and R.I. determinations (table 2) suggest that it is a fairly diopsidic augite. It is usually faintly tinted green or brown and is often altered to chlorite with calcite (R901) or quartz (R920) rarely present as well.

*Chlorite* occurs as well defined fragments or as rather irregular structureless patches

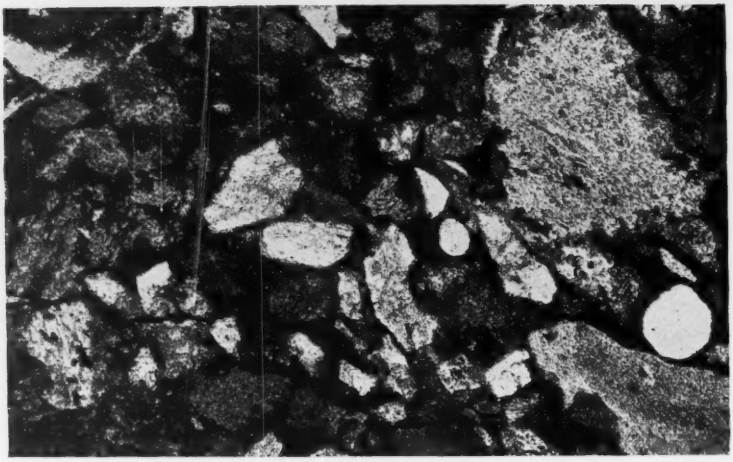

FIG. 6.—Lithic labile graywacke (R924) from Yarrimie Formation, showing prominent matrix (×55).

The latter are of a shape which could not have withstood transport and presumably result from alteration in place of other material such as glass.

In general the well defined grains are decussate aggregates and lack inclusions. The irregular patches are usually structureless but occasionally show banding reminiscent of flow structure. Inclusions of plagioclase, epidote, clinozoisite, apatite, and opaques are frequent. In R904 the chlorite is pleochroic and apparently pseudomorphous after biotite.

The color and optical properties of the chlorite vary widely, four different types being found in R904. In most slides the chlorite is some shade of yellow-green. Bright green and blue-green varieties also occur. Signs of elongation indicate that the chlorites are optically negative.

*Calcite* is never abundant. Occasional crinoid and brachiopod fragments are seen, but often the derivation of the detrital calcite is not apparent.

*Rock fragments* are dominantly volcanic, many varieties occurring. Frequently they show no signs of abrasion, boundaries being very irregular (R909). In most cases the feldspar in the volcanics is albite, but in R907, R909, and R920 more calcic feldspar is present, at times veined by albite. In such cases the fragments can be identified as andesite. In general pyroxene is restricted to unaltered rock fragments (R907, R909). In the albitic volcanic fragments ferromagnesians are replaced by decussate masses of chlorite which may also form vesicle fillings. Calcite and quartz-albite may also occur in vesicles. Glassy volcanics are always devitrified and are abundant in R915 and R921. The textural varieties of volcanics encountered are listed in table 3.

In most specimens there is an acid volcanic component consisting of fine, holocrystalline feldspathic types, mostly rhyolites, with interstitial primary quartz.

Granite fragments occur occasionally (R904, R907, R918, R920) but are always a minor constituent. Micrographic granite occurs in R907.

Siltstone fragments, which frequently contain radiolaria, form a minor constituent of some rocks. In R905 the radiolaria may be replaced by prehnite. These siltstones are similar to those interbedded with the arenites and are probably derived therefrom. Fragments of vitric tuff and occasional shards, all now devitrified, occur in R924. Limestone fragments occur in R905 and R907. R903 contains a quartz-epidote-opaque-chlorite rock of unknown origin.

*Accessories* include opaques of irregular

TABLE 3.—*Varieties of rock fragments in detritus*

| Stratigraphic Unit | intergranular | pilotaxitic | trachytic | subtrachytic | arabesquitic | vitric | felty | subophitic | hyalopilitic | pseudoclathrate | perlitic | fluidal | spherulitic | pumiceous | hyaloophitic | textural andesite | textural basalt | opaque-rich | subpilotaxitic | vesicular | granular |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Dmy (lithic lab. gwkes) | tr | x | p, v | p | x | o, b | x | x | p, v | x | x | x | v | x | x | x | x | x | | x | |
| Dmlg | | | x | | x | v | x | x | | x | | | | x | | x | x | | x | x | |
| Dms | | | x | | | o, b, v | x | | x | | x | x | | | | | x | | | x | |
| Dld (lithofeldspathic) | | | x | h | | | x | x | | | | | | x | h | | | | | | |
| Dld (hematitised) | h | h | x, h | h | x | | h | | x | | | | | x, h | h | | h | | | h | h |
| Dh | | | x | | | x | x | | | | | x | | | | x | x | | | | |

tr.—trachytic; p.—porphyritic; v.—vesicular; h.—hematitised; o.—orange; b.—brown; x.—present.

shape, green epidote, sphene, apatite, amphibole (pleochroic from pale yellow-brown to deep brownish-green) (R925), and sphene-chlorite fragments consisting of an octahedral network of sphene lamellae with chlorite occupying the interstices.

*Matrix*, when prominent, is grayish brown apparently largely chloritic, and shows little sign of reconstitution (fig. 6).

*Cement* consists of one or more of chlorite, carbonate, prehnite, quartz, albite, epidote, pyrite, and ?pumpellyite. The percentage of cement varies inversely with the percentage of matrix and is minor volumetrically. It shows no signs of distortion and therefore post-dates the extreme compaction of the sediments.

*Veins* are occasionally encountered. R909 contains chlorite-carbonate-quartz and also epidote; R919, albite-carbonate; R921, quartz-carbonate-albite-prehnite, quartz-feldspar-muscovite-carbonate and prehnite; and R924 contains a chlorite-carbonate vein.

*Feldspathic labile arenites.*—These are coarse and well sorted, consisting largely of plagioclase and quartz. Matrix is dominant only in R914, where a roughly fluidal structure is present. Fragments, particularly quartz, are well rounded but of variable sphericity. Rock fragments tend to be less well rounded. Broken grains are common except in R914.

Of the three specimens examined, R913 and R914 are rather altered. In the former the feldspar is entirely pseudomorphed by aggregates of strongly absorbing carbonate; in the latter, alteration makes it difficult to tell whether one is dealing with a volcanic or a sediment.

*Quartz* in R913 is simple and very strongly strained, occasionally simulating perthite. Embayments, infilled with a fine quartz mosaic, are frequent. Grains may be clouded with fluid inclusions which are often in trails, and zircon inclusions have been noted. Carbonate replaces quartz along fractures.

In R914 the quartz is simple, unstrained, and embayed, with abundant fractures which may contain muscovite. Acicular, unidentifiable high relief, and fluid inclusions, often in trails, occur. Embayments may be filled by sheaf-like chlorite masses. The quartz in R910 is similar but much less frequently embayed. Some composite non-sutured grains occur.

*Feldspar* is completely replaced in R913. In R914 it is albite, but is almost completely replaced by carbonate and sericitic and limonitic material. In R910 the feldspar is fairly calcic, multiple twinned, zoned, veined by albite, and locally clouded by sericite. Apatite inclusions occur.

*Accessories* comprise leached biotite and apatite (R914), opaques, calcite, chlorite—as structureless patches with opaque inclusions—and rock fragments (R910). The last have andesitic affinities and include pilotaxitic, trachytic, porphyritic hyalopilitic, and vesicular vitric varieties.

*Matrix* in R914 is extensively stained with limonite and is finely aggregate-polarising.

*Cement* is usually a quartz mosaic, joined by pyrite in R910. The mosaic may heal fractures in the quartz grains. In some cases the mosaic appears to have attacked the quartz grains, which may be the origin of some embayments. A quartz-carbonate vein occurs in R914.

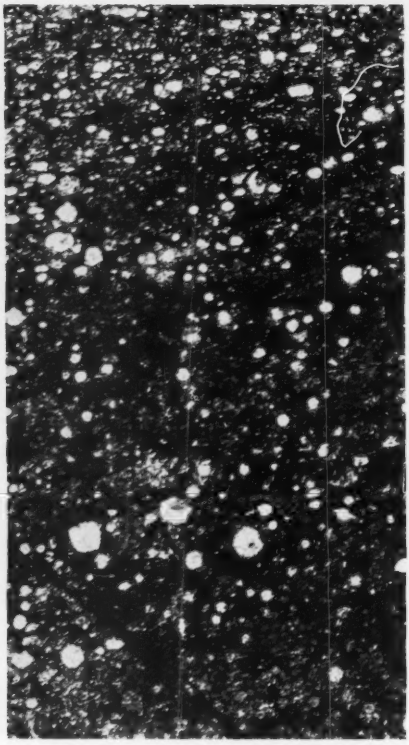

FIG. 7.—Black and white banded radiolarian argillite (R908) from Yarrimie Formation, section of black band (×22).

*The argillites.*—(R908, R909, R911, R912, R916, R953) None of these can be termed chert because silty detritus is always too abundant. All contain numerous radiolaria, and it is probable that their $SiO_2$-contents are well above the average for shales.

The black and white banded variety contains two distinct components. The black bands consist of very fine laminae (up to 0.07 mm thick) of dark opaque material with chips of quartz, muscovite, and minor albite and abundant radiolaria (fig. 7). The radiolaria are recrystallised and replaced by a mosaic of quartz and chlorite, occasionally with some calcite and chlorite (R908).

The white bands are siltstone or fine arenite and may show poor graded bedding. They consist almost exclusively of closely packed grains of albite, heavily altered in R916 to sericite and carbonate. A little yellow-green chlorite, fine rock fragments, quartz, and epidote complete the assemblage. Matrix is often present, but its place in R916 is taken by a quartz-albite cement. A prehnite vein was noted in R908.

The green argillites are fine siltstones, usually with some lamination, which are dominantly chloritic with scattered chips of feldspar, quartz, and opaques (fig. 8). Muscovite is always present, and epidote and amphibole occur in R909. Bright green chlorite filaments up to 0.02 mm x 0.2 mm occur in R912. Fine aggregate-polarising material is usually abundant.

Radiolarian tests up to 0.2 mm diameter are abundant in the green argillites. These are generally recrystallised to a quartz mosaic, often with a central mass of chlorite. Replacement by prehnite occurs in R909 and R953. Not infrequently structures are preserved. *Stylosphaera sp.* is abundant in R911 (fig. 8). Sponge spicules, also converted to quartz mosaics, may accompany the radiolaria (R912). They appear to be more abundant in the slightly coarser bands.

In R953 brecciation of the argillite has occurred. The fractures are healed by aggregate polarising quartz, probably inverted from opal. Later quartz-prehnite-muscovite veins transect the rock.

*The argillaceous limestones (R903, R906).*—These are fine, very rich in calcite, which is apparently of chemical origin, and contain radiolaria and appreciable quantities of silt. Detrital quartz, feldspar, and opaques are usually present. In addition R906 contains pale green chlorite pseudomorphs after glass. These show vesicular structure and contain feldspar laths (fig. 9); secondary pyrite and partial replacement by calcite occur. Volcanic fragments—textural andesite, and trachytic and vesicular vitric types—also occur in R906. Diagenetic prehnite occurs in R906 and chlorite in R903.

### Levy Graywacke Member (Dmlg)

(R926, R927) This consists of very coarse lithic graywacke, locally rudaceous, which is very poorly sorted and has little matrix. Petrographically it is similar to other Yarrimie Formation lithic labile graywackes, but

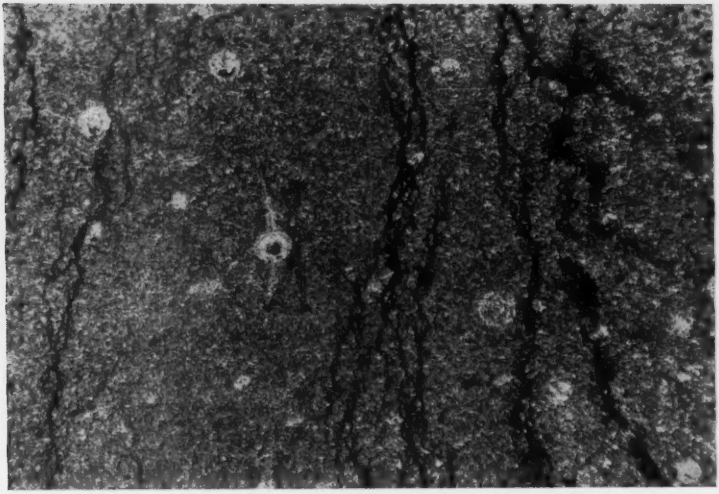

FIG. 8.—Dark green argillite (R911) with scattered radiolaria (*Stylosphaera sp.* near center), from Yarrimie Formation (×44).

its rock fragments tend to be rather reconstituted, calcite, chorite, and albite being prominent.

The detritus has few notable features. Quartz is simple and unstrained; amphibole is common and is partly replaced by chlorite. The textural varieties of rock fragments encountered are listed in table 3. Dolerite and microgranite fragments also occur.

### Silver Gully Formation (Dms)

(R928–32) This formation is dominantly rudaceous, but appreciable quantities of arenite occur. All specimens examined are

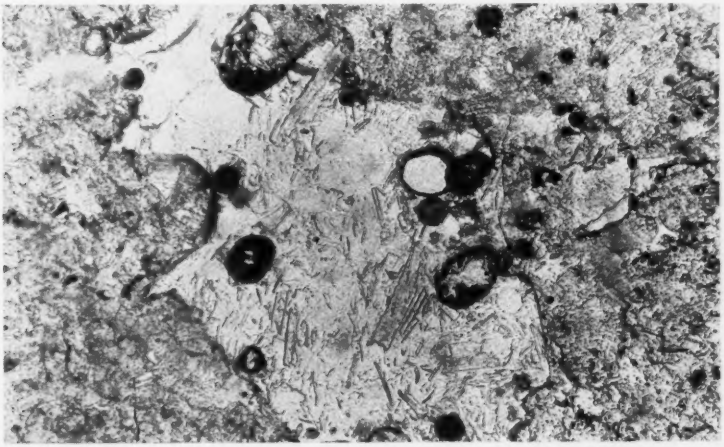

FIG. 9.—Fragment of glass, now pseudomorphed by chlorite, containing feldspar laths (×110). R906, an argillaceous limestone from the Yarrimie Formation.

lithic labile graywackes with the exception of R931, a fine tuffaceous graywacke, and are usually medium to coarse arenites or fine rudites. They are generally moderately to well sorted with minor matrix and cement. Roundness is rather variable, the smaller fragments being angular, the larger rounded. Sphericity is fairly constant at about 0.7. Fragments are in close contact, indicating a high degree of compaction. The rocks are well indurated and dominantly gray with the colors of the individual minerals apparent. Occasionally bright green is dominant, apparently resulting from an abundant chlorite cement.

The graywackes consist dominantly of rock fragments, chiefly volcanic, and abundant feldspar, either albite or andesine. Minor quartz, chorite, pyroxene, amphibole, carbonate, sphene-chlorite, opaques, and apatite complete the assemblage.

R932 shows signs of metamorphism. The matrix is reconstituted and abundant needles of actinolite occur both between grains and within rock fragments. Grain boundaries are still sharply defined, and the texture of the volcanic fragments often remains unaltered.

*Quartz* is present only in R928 and R929 and consists of simple unstrained grains with scattered fluid inclusions.

*Feldspar* is commonly albite ($An_2$) similar to that in the Yarrimie Formation. Andesine ($An_{42}$) occurs in R929 and R930, in the former being veined by albite ($An_2$). It is clear, multiple twinned, at times zoned, and contains apatite inclusions. Locally it may be sericitised or replaced by carbonate.

*Ferromagnesians*, both pyroxene and amphibole, are present. The former is similar to that in the Yarrimie Formation and is at times replaced by carbonate. The amphibole is pleochroic from yellow-brown to blue-green and occurs only in R932.

*Chlorite* is generally yellow- or blue-green and occurs as irregular patches of the type already described. It is very abundant in R931 where it is apparently pseudomorphous after glass and also in the volcanic fragments of R929 and R930 where it imparts a distinct green color to the rock. It appears to pass to amphibole in R932.

*Rock Fragments* are again dominantly volcanic, but the total aspect is not as obviously andesitic as in the Yarrimie Formation. Alteration to albite-chlorite-calcite with the preservation of original textures is frequent but the original more calcic plagioclase is preserved in R929 and R930. In addition to the textural varieties listed in table 3, basalt, altered dolerite, and felty and trachytic andesite have been recognised. The textural basalt (table 3) may contain pyroxene.

Occasional fragments of graywacke?-siltstone, limestone, and possible metaquartzite occur. Granite occurs in the rudites. Considerable quantities of isotropic fragments, probably glass, occur in R930. R931 is composed largely of altered pumiceous glass, but shards have not been noted.

*Accessories* listed above are similar to those of the Yarrimie Formation. The carbonate is both organic and inorganic. An aggregate of prehnite, which may be diagenetic, occurs in R929.

*Cement* is not prominent; quartz, albite, and chlorite occur, the chlorite in R930 being moulded around the grains. Diagenetic modification of glass is frequent, that in R931 being replaced by chlorite and prehnite. Perlitic fragments in other slides now consist of albite aggregates with chlorite tracing out the perlitic fractures (fig. 10).

### Wogarda Argillite (Dlw)

This unit consists dominantly of pale gray siliceous argillites which are often silty. They contain occasional thin arenite interbeds with graded bedding; one of these (R933) is a feldspathic labile graywacke.

The rock is very well sorted, medium grained, with fragments mainly subangular or subrounded of sphericity about 0.4. Compaction is not as intense as in higher formations. Quartz, feldspar, and opaques together with minor chlorite and rock fragments comprise the bulk of the detritus. Cement and a somewhat reconstituted matrix are present.

*Quartz* grains are simple and unstrained or slightly strained. Some are well rounded. Fluid inclusions, isolated and in trails, zircon, and rod-shaped inclusions have been noted.

*Feldspar* is very heavily kaolinised and at times shows multiple or chequer-board twinning. Apatite inclusions occur. Probably both orthoclase and albite are present, but

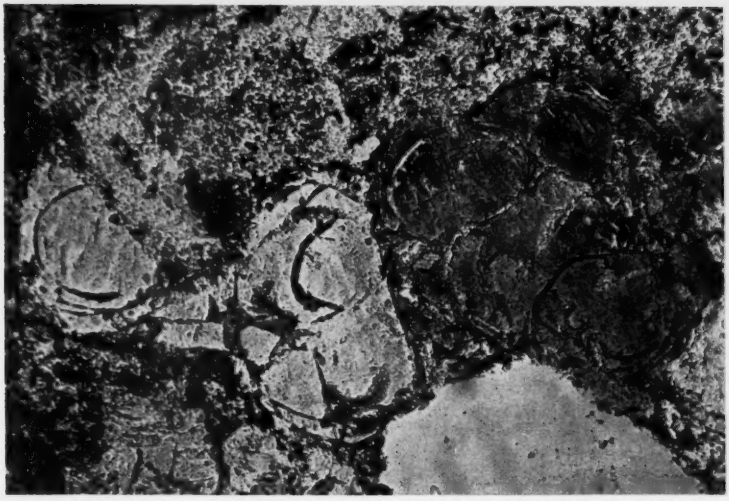

FIG. 10.—Volcanic fragment showing perlitic structure (×72). The glass is replaced by albite and the perlitic cracks are traced out by chlorite. R928, a lithic labile graywacke from the Silver Gully Formation.

the kaolinisation prevents certain determination.

*Accessories* comprise optically negative yellow-green chlorite forming irregular patches, minor amounts of granite and felty volcanics of obscure affinities, epidote, apatite, and opaques, some of which contain apatite.

*Matrix* consists of a chlorite-quartz-albite mosaic. It is reconstituted and passes into the cement which consists of the same minerals. There has been secondary enlargement of the quartz and reconstitution of some volcanic fragments. An albite-carbonate vein cuts the rock.

### Drik-Drik Formation (Dld)

The upper portion of this unit consists mainly of green arenites (R937, R948) which contain large amounts of feldspar and rock fragments. Both lithofeldspathic and feldspatholithic graywackes are probably represented; graded bedding occurs in various places.

The graywackes are generally coarse and pebbly, some bordering on rudites. Grains range from angular to subrounded, most being subangular. Sphericity averages 0.6 but is variable. Strong compaction has occurred, and the matrix, which is fairly prominent, is somewhat reconstituted.

*Quartz* grains are generally simple and strained, show embayments, and may contain fluid and high relief inclusions.

*Feldspar* is either extensively sericitised albite ($An_2$), or labradorite ($An_{52}$) extensively veined by albite ($An_2$). It is rarely multiple twinned and may contain calcite filling fractures.

*Rock Fragments* are dominantly igneous and include the varieties listed in table 3. All are rich in albite and chlorite, the pumice being completely replaced by pink albite. Some are almost opaque because of hematite; these are red in hand specimen and are characteristic of these graywackes as some grains appear in every hand specimen. The hematitised fragments are similar to the dominant constituent of the red rudites lower in the formation.

Sedimentary fragments, which are minor, comprise vitric tuff with authigenic albite, graywacke?-siltstone, and feldspathic labile arenite with abundant chlorite and pink albite. Glass shards, replaced by pink albite, occur in R948.

*Accessories* consist of chlorite in irregular structureless patches with feldspar and opaque inclusions, inorganic calcite, opaques, epidote, apatite, and a little pyroxene partly replaced by calcite (R948).

*Diagenesis* has produced a reconstituted albite-chlorite matrix, and there is no cement. Diagenetic modification of rock fragments is widespread with the production of chlorite, albite, and possible pumpellyite. The latter is finely granular, of high R.I., D.R. about 0.015, and pleochroic from blue-green to golden brown. Strong dispersion prevents complete extinction. Quartz-albite-?laumontite-chlorite veins occur in R937.

The lower part of the formation consists mainly of purple-red rudites which are often porous and some red arenites. The rocks are cemented by pink and white feldspars with a little green epidote and chlorite; this produces a striking appearance in hand specimen. The rudites (R938–40, R943–47) are well sorted and consist almost entirely of strongly hematitised rock fragments.

*Rock Fragments* are the only detrital constituent observed in these rudites. The fragments are frequently opaque because of hematite and often contain laths of albite which may show traces of multiple twinning. The shape of the fragments is usually extremely irregular, and it is unlikely that they have been subjected to appreciable transport. Vesicular varieties show development of the minerals of the cement within the vesicles. The textural varieties encountered are listed in table 3. Dolerite fragments also occur.

Some fragments are not hematitised; they may be extensively chloritised and some show vesicular structure. With increasing chloritisation they pass into irregular patches of blue-green chlorite, sometimes with the vesicular structure preserved.

*Cement* consists dominantly of colorless and pink albite ($An_2$), orthoclase, epidote, and chlorite, with some pink or colorless zeolite (probably laumontite), calcite, and quartz. In some cases the distribution of the hematite is suggestive of its being a cement. Calcite-laumontite veins occur in R947.

The red arenites (R935, R941, R942) contain appreciable calcite in the form of organic debris and oölites. The internal structures of the oölites are often traced out by hematite, and a nucleus of laumontite is occasionally present (R935). Some grains appear to be of algal origin; structures similar to *Pachytheca* Hooker and *Palaeoporella* Stolley have been noted (figs. 11 & 12). Ostracod shells also occur (fig. 12).

A little detrital quartz and albite occur. This and the calcareous detritus may be set in a mass of hematite (R935) or calcite-epidote-chlorite (R941). Calcite in R935 is anomalously biaxial: $2V(-)$ ca. 15°, $n_Y = 1.657$.

A similar type, now metamorphosed with the development of muscovite and epidote (R936), is closely associated with R935 in the field. It is also red.

### Pipeclay Creek Formation (Dlp)

Argillites are dominant in this formation, both black and white banded and green cherty types being present. Occasional interbeds of graywacke occur. The black and white argillites are similar to those of the Yarrimie Formation. The black bands have laminae 0.01–0.04 mm thick and are rich in dark brown material but lack radiolaria. The white bands are noticably less feldspathic than their counterparts in the Yarrimie Formation but contain chips of quartz and albite set together with radiolaria in a fine aggregate-polarising mass of chlorite and possible devitrified glass (R950).

The green argillites (R949) are less conspicuously banded and contain appreciable fine silt and scattered radiolaria which may be replaced by prehnite. Slivers of quartz, albite, muscovite, and granules of epidote and chlorite have been observed. The finer material is aggregate-polarising and irresolvable and contains patches of calcite and prehnite, the latter being more dominant in the paler bands.

One lithic labile graywacke has been examined (R951). It is somewhat metamorphosed with the matrix reconstituted to actinolite and biotite in needles and poikiloblastic plates respectively. Quartz occurs as strained simple grains with occasional fluid and acicular inclusions. The plagioclase is always fairly calcic with sharp multiple twinning, zoning, and occasional apatite inclusions. It is often clear but some is heavily clouded. Altered radiolarian siltstone and relics of opaques occur rarely.

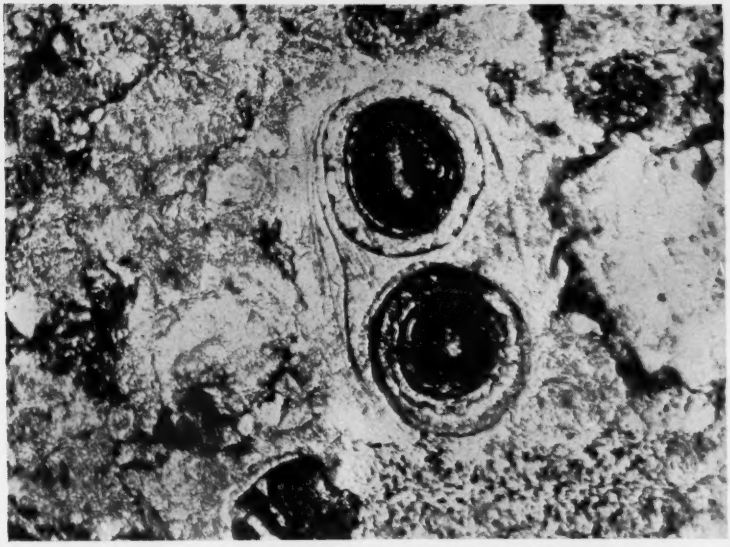

FIG. 11.—A calcareous alga (?), with structures traced out by hematite (×94). R 941, a calc-arenite from the Drik-Drik Formation.

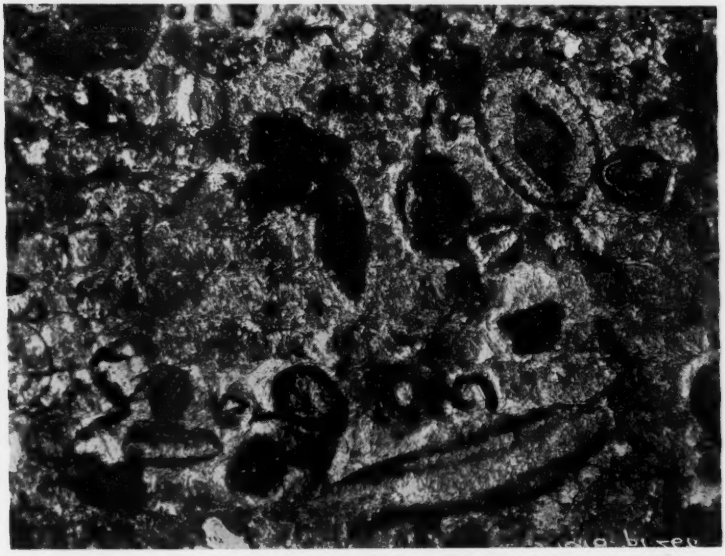

FIG. 12.—Ostracodes, and possible calcareous alga (*?Pachytheca*) in upper right corner (×61). Same section as figure 10.

### Seven Mile Formation (Dls)

This unit, consisting of argillites with some graywackes, is metamorphosed by the Moonbi Granite. One of the less altered arenites (R952) was almost certainly a lithic labile graywacke with some calcite, although no definite rock fragments remain. A hornfelsic structure is developed with abundant pyroxene and some wollastonite. Relics of clastic feldspar, probably albite, remain.

### Hawk's Nest Beds (Dh)

This unit consists of black fissile argillites with interbedded graywackes. At times the graywackes are metamorphosed by the Mount Ephraim Granite with the development of actinolite and biotite (R957, R958).

Both feldspathic and lithic labile graywackes occur. The former (R955, R959) are medium to fine grained, very well sorted, and strongly compacted. Fragments are subangular to subrounded and have a sphericity averaging 0.7. Quartz and plagioclase are the major components, and cement is prominent. They are similar to the feldspathic labile graywackes of the Yarrimie Formation but lack prominent matrix.

*Quartz* occurs as simple unstrained grains with high relief and fluid inclusions, often in trails.

*Feldspar* is dominantly moderately calcic plagioclase, extensively veined by albite or a zeolite, probably laumontite (R915). The feldspar is usually clear and may show multiple twinning. There is also a small amount of albite which is flecked with sericite and partly replaced by calcite. Some show multiple twinning.

*Rock Fragments* occur only in R959 and consist of quartzite, textural basalt, cryptofelsitic and felty volcanics, and scattered siltstone chips.

*Accessories* consist of yellow- or blue-green optically negative chlorite in structureless patches, biotite, apatite, opaques, and calcite which is in part organic.

*Diagenesis* has produced a chlorite-zeolite (?laumontite) cement in R955 and a calcite cement in R959.

The lithic labile graywackes (R957, R958, R960) are coarse, some bordering on fine rudites, and poorly sorted. Fragments are generally subangular to subrounded and sphericity averages 0.7. Compaction is strong and matrix is usually fairly prominent.

*Quartz* is a minor constituent and occurs as simple unstrained grains with fluid, rod, and high relief inclusions. Orthoclase may be associated with the grains.

*Feldspar* is either albite similar to that elsewhere in the sequence or moderately calcic plagioclase extensively veined with albite which is locally replaced by calcite.

*Rock Fragments* are dominantly volcanic, textural varieties being recorded in table 3. All are somewhat altered and contain albite. Radiolarian argillite fragments occur in R957, granite in R960, and probable vitric tuff in R958.

*Accessories* include apatite, calcite (in part organic), and opaques. Chlorite occurs as blue-green decussate grains in R960, and amphibole pleochroic from yellow-brown to blue-green is abundant in R958.

*Matrix*, which is prominent, is metamorphosed to fine biotite and acicular actinolite in R957 and R958.

*Diagenesis* has produced a chlorite-calcite cement in R960. No cement remains in other slides. R957 contains a quartz-bytownite ($An_{81}$) vein with accessory sphene and apatite. An albite-calcite-actinolite vein occurs in R958.

A polymictic granule rudite (R956) was also examined. It consists of subangular to subrounded grains of very variable sphericity cemented by coarsely crystalline calcite. The cement is early diagenetic; the small number of grain contacts visible indicates a very low degree of compaction.

Detritus consists of minor amounts of strained quartz, pale yellow-brown pyroxene partly replaced by chlorite and calcite, albite with chlorite inclusions, and rock fragments. The latter are dominant and comprise limestone; siltstone; pilotaxitic, arabesquitic, porphyritic felty, and other altered volcanics; and patches of chlorite with calcite inclusions which were probably originally glass.

## DISCUSSION

### Lithic Labile Graywackes

The lithic labile graywackes in the Yarrimie Formation, which are similar to those of the overlying Parry Group, are domi-

nantly of andesitic derivation. The rock fragments are texturally andesitic types, and their feldspar, together with the free feldspar grains, was originally andesine-labradorite. The pyroxenes, amphiboles, and opaques, which occur both in the rock fragments and as free grains, are likewise characteristically andesitic.

Some of the quartz is volcanic and, with the rhyolite fragments, forms an acid volcanic component. This is at times accompanied by plutonic quartz, orthoclase, and granite forming an acid plutonic component. Sedimentary fragments are largely autochthonous with the exception of the vitric tuffs.

The lithic labile graywackes and related rudites lower in the sequence (excepting the lower part of the Drik-Drik Formation) are similar to those above but differ in the textural varieties of volcanic fragments which are present. In the Silver Gully Formation andesitic detritus is again dominant and is accompanied by many devitrified glassy volcanic fragments. The acid component is very minor, being solely granitic. Allochthonous material is present in the sedimentary component which is stronger than in the Yarrimie Formation. Autochthonous tuffaceous material is also present.

In the upper part of the Drik-Drik Formation the andesitic component, although present, is not as clearly recognisable. It is accompanied by keratophyre, vitric tuff, and a little dolerite. The sedimentary component is again allochthonous. Embayed quartz suggests an acid volcanic component.

The Hawk's Nest Beds, whose stratigraphic position is unknown, contain detritus similar to that in the Yarrimie Formation.

### Feldspathic Labile Arenites

These are rather variable but appear to contain two major components—andesine, presumably derived from andesitic crystal tuffs, and quartz of unknown derivation. The presence of granite, strained quartz, or allochthonous sedimentary material in some slides suggests that the quartz may be dominantly plutonic. A minor lithic component of andesitic derivation is not uncommon.

### Radiolarian Argillites

The black and white radiolarian argillites of the Yarrimie Formation resemble the diatomaceous sediments of the Monterey Formation of California (Bramlette, 1946). The main points of similarity are:

1.—High $SiO_2$-content of the argillites
2.—Extensive thin sedimentation units, showing fine lamination and some graded bedding
3.—Argillaceous limestones, of diagenetic chemical origin, around which the argillites are flexed because of differential compaction.

Three analyses of Yarrimie Formation argillites (David and Pittman, 1899) give $SiO_2$—67.87%, 80.5%, and 91.6%. Normal shales average 58% $SiO_2$ (Pettijohn, 1957, p. 106); the lowest silica-content of several cherts given by Pettijohn (1957, p. 435) is 85%. Of the three analysed argillites the last is a chert and the others are siliceous argillites. The range of silica contents of Monterey Formation rocks given by Bramlette (1946, p. 13) as 71.8%–92.37% is comparable with that of the Yarrimie Formation.

The rhythmic alternation of bands rich in terrigineous detritus and bands rich in biogenic material is characteristic of both the Yarrimie and Monterey Formations. However, larger and smaller scale rhythms superimposed on this rhythm which are recorded from the Monterey Formation (Bramlette, 1946) have not been observed in the Yarrimie Formation.

### Argillaceous Limestones

The detrital components of these limestones are closely similar to those of the enclosing argillites and the graywackes. Tuffaceous material is more easily recognisable because the outlines of replaced glass fragments have not been destroyed by compaction.

The rocks are closely similar in occurrence and probably in origin to the carbonate concretions described by Bramlette (1946, p. 20) from the Monterey Formation of California. Stratification is continuous from the argillites into the limestones, and the detritus is similar. The calcite was probably introduced during early diagenesis and has prevented the intense compaction of the

rocks. Like their Monterey Formation counterparts these rocks contain excellently preserved microfossils.

### Red Sediments of the Drik-Drik Formation

These rocks, which are distinctly different from anything elsewhere in the sequence, appear to be entirely of volcanic derivation except for the oolitic and biogenic detritus in the fine arenites. The volcanic material appears to be largely of keratophyric origin, but determination is made difficult by the intense hematitisation.

It is probably significant that this unit immediately overlies the Cope's Creek Keratophyre, and possibly grades into it (Benson, 1915a, p. 155). The shapes of the volcanic fragments indicate a very short transportation history, and it may be surmised that they have been derived either from the Cope's Creek Keratophyre or from volcanics contemporaneous with it situated on a nearby land mass.

The hematitisation is probably due to subaerial weathering of the detritus prior to deposition. Its preservation suggests either terrestrial deposition, or rapid deposition under marine conditions. The presence of oolites indicates that at least part of the red sediments are marine and probably shallow-water deposits.

## NATURE OF SOURCE AREAS

Throughout the period of deposition of the Tamworth Group source areas were dominantly composed of volcanics. These were initially keratophyres which gave rise to the earlier Drik-Drik Formation sediments, but subsequently andesites became dominant, probably appearing first during the closing stages of deposition of the Drik-Drik Formation.

The Wogarda Argillite represents an interlude during which little coarse detritus reached the trough, but occasional feldspathic labile graywacke beds indicate a continuation of vulcanicity in surrounding areas.

With the commencement of deposition of the Silver Gully Formation the dominance of andesite in the source areas becomes obvious. This continues into the Yarrimie Formation, which represents a period of intermittent deposition of coarse detritus in the trough and upwards into the overlying Parry Group.

Old sediments appear to have formed a small part of source areas from late Drik-Drik Formation times until the close of deposition of the Silver Gully Formation. They do not appear to have contributed to the detritus which formed the Yarrimie Formation. Granite however appeared in the source areas during deposition of the Silver Gully Formation and continued to contribute to the detritus during the formation of the Yarrimie Formation.

Acid volcanics may have been present in the source areas during formation of the later Drik-Drik Formation sediments. They were definitely present during deposition of the Yarrimie Formation, but it is not known whether they appeared as a result of contemporaneous acid vulcanism, or as a result of the exposure of old acid volcanics by erosion.

The dominantly andesitic nature of the detritus throughout most of the sequence and the deep-water deposition of the sediments strongly suggest that the source areas consisted of an island arc or arcs. These were probably composed of a folded sedimentary belt intruded by granite and almost completely covered by andesitic debris, chiefly in the form of crystal-lithic tuff cones and sheets, with continual addition of more andesitic material.

## ACKNOWLEDGEMENTS

The author gratefully acknowledges financial assistance from a University of New England Research Grant, the assistance of Mr. N. Petrasz in the preparation of slides, and of Mr. G. Taylor in the photography.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 370–379
FIGS. 1–2, SEPTEMBER, 1960

# OCEAN BASIN AGES AND AMOUNTS OF ORIGINAL SEDIMENTS[1]

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## ABSTRACT

The present sediment thicknesses in deep-sea basins, away from continents and islands, are apt to be the total thicknesses of the first two (or more) layers revealed by seismic surveys. The application of soil mechanics techniques to this and associated problems indicates the possibility of determining approximate amounts of original sediments necessary to consolidate and lithify into present thicknesses. Once this is done, then computations can be made of the amounts of solids in the original sediments, and, assuming rates of sedimentation, estimates can be made of the ages of ocean basins under specific localities. In general, these computations and estimates indicate sufficient solids are present on the sea floor to satisfy the needs of the world-wide geochemical balance and that the ocean basins are of ancient age (Paleozoic to Pre-Paleozoic).

## INTRODUCTION

Seismic surveys in all of the ocean basins have revealed, normally, a thin cover of material of low velocity considered to be "unconsolidated sediment." These surface sediments are usually underlain by a second layer whose elastic-wave transmission velocities are intermediate between those of the sediment and the basaltic rock of the oceanic crust.

In a previous paper (Hamilton, 1959) the writer discussed at some length the probability that in many areas the "second layer" is consolidated and lithified material of the type now being deposited in the deep-ocean basins (mainly clay and calcareous ooze). The evidence lies mainly in the fields of soil mechanics and geophysics and will not be repeated here except to summarize for background for the present paper.

In general, gravitational consolidation of both clay and calcareous ooze should result in decreases in porosity and pressure-chemical effects so that lithification into shale and limestone should take place at depths in the sediment between 150 and 700 meters with marked increases in seismic velocity at these depths. The present sediment thickness in the deep-sea basins away from continents and islands is apt to be the total thickness of the first two layers. If so, the thickness of the second layer should enter into the calculation of total sediment deposited in the ocean basins, and the matter assumes importance in the geochemical balance.

The present paper will deal with computations and discussion of the original amounts of sedimentary material necessary to form the present sedimentary column overlying the basalt of the ocean basins at several specific localities, assuming that the thickness is made up of "layers one and two" (or more) of seismic literature. Once this is done, computations can be made of the amount of solids in the original sediments, and assuming rates of sedimentation, one can estimate ages of the basins under specific seismic stations.

## AMOUNTS OF ORIGINAL SEDIMENTS

### General

In order to predict settlement or consolidation of a sediment under a load (either due to its own overburden weight or a new load such as a building) all that is necessary is an estimate or measurement of the void ratio, $e$ (ratio of the volume of voids to volume of solids), before and after imposition of the load. In soil-mechanics engineering practice a sample of sediment is usually subjected to a consolidometer test and a curve drawn showing the decrease of the void ratio under pressure. When the expected new load is computed, the curve can be entered along the pressure line and the new or final, void ratio determined. After many years of use, it has been determined that this method

[1] Manuscript received November 30, 1959.

TABLE 1.—*Consolidation of clay and shale (from Hamilton, 1959)*

| Depth in Sediment (m) | Interval Pressure ($Kg/cm^2$) | Cumulative Pressure ($Km/cm^2$) | Void Ratio, e | Density sat. (gr/cc) | Submerged wt. (gr/cc) | Porosity, n (%) |
|---|---|---|---|---|---|---|
| 0 | 0 | 0 | 2.58 | 1.40 | 0.35 | 72 |
| 10 | 0.35 | 0.35 | 1.73 | 1.52 | 0.47 | 63 |
| 20 | 0.48 | 0.83 | 1.48 | 1.55 | 0.50 | 60 |
| 30 | 0.51 | 1.34 | 1.35 | 1.58 | 0.53 | 58 |
| 40 | 0.54 | 1.88 | 1.25 | 1.60 | 0.55 | 56 |
| 50 | 0.56 | 2.44 | 1.18 | 1.63 | 0.58 | 54 |
| 60 | 0.59 | 3.03 | 1.13 | 1.64 | 0.59 | 53 |
| 70 | 0.60 | 3.63 | 1.08 | 1.66 | 0.61 | 52 |
| 80 | 0.62 | 4.25 | 1.05 | 1.67 | 0.62 | 51 |
| 90 | 0.63 | 4.88 | 1.02 | 1.68 | 0.63 | 50 |
| 100 | 0.63 | 5.5 | 0.98 | 1.69 | 0.64 | 49 |
| 150 | 3.3 | 8.8 | 0.87 | 1.72 | 0.67 | 47 |
| 200 | 3.5 | 12.3 | 0.80 | 1.76 | 0.71 | 44 |
| 250 | 3.6 | 15.9 | 0.75 | 1.77 | 0.72 | 43 |
| 300 | 3.7 | 19.6 | 0.70 | 1.80 | 0.75 | 41 |
| 400 | 7.7 | 27.2 | 0.58 | 1.85 | 0.80 | 37 |
| 500 | 8.1 | 35.3 | 0.54 | 1.87 | 0.82 | 35 |
| 600 | 8.3 | 43.6 | 0.50 | 1.90 | 0.85 | 33 |
| 700 | 8.6 | 52.2 | 0.47 | 1.91 | 0.86 | 32 |
| 800 | 8.7 | 60.9 | 0.43 | 1.93 | 0.88 | 30 |
| 900 | 8.9 | 69.8 | 0.40 | 1.95 | 0.90 | 29 |
| 1000 | 9.1 | 78.9 | 0.37 | 1.97 | 0.92 | 27 |
| 1500 | 48.4 | 127.3 | 0.25 | 2.06 | 1.01 | 20 |
| 2000 | 52.8 | 180.1 | 0.15 | 2.15 | 1.10 | 13 |

Assumption: 300 m of clay overlie shale with 100 percent hydrostatic uplift effective. Properties: Density of solids: 2.31 gr/cc; water density: 1.05 gr/cc; initial porosity: 72 percent.

yields valid results both in construction of buildings and in predicting gravitational consolidation.

Almost all clay and silty clay sediments, regardless of original porosity, will reach an approximate common porosity (approximately 29 percent) when the overburden pressure reaches 100 $kg/cm^2$ (Skempton, 1953; Hamilton, 1959). This phenomenon allows the use of generalizations in studying the effects of overburden pressures on consolidation even though the exact sediment type (in the silt-clay range) and void ratios at the site are actually unknown.

Assuming properties of the surface sediments, it is then possible to construct a table (table 1) showing the variations of properties with depth in the sediment by using standard soil mechanics techniques and the cumulative intergranular buoyed weight of the mineral grains as induced pressure. These matters are discussed at some length in a previous paper (Hamilton, 1959) and can be found in most soil mechanics texts.

A calcareous ooze will consolidate along curves typical of sandy silt or sand-silt-clay and will always be on the high porosity side of similar curves for clay (Hamilton, 1959).

Studies of shale sections on land, seismic work at sea, and theoretical studies indicate that clay should become shale (due to pressure-chemical effects) at depths between 150 and 700 meters with a probability that the depth will be near 300 meters (Nafe and Drake, 1957; Hedberg, 1936; Hamilton, 1959). Inasmuch as calcareous material is more susceptible than clay to lithification, this material should become limestone at similar or less depths.

When the pressure-void ratio relationships for any sediment are known (or estimated) it is possible to determine the approximate amount of settlement, or consolidation, to be expected under any given load. Conversely, it is also possible to determine how much sediment was required to consolidate to present thicknesses. These are standard techniques in soil mechanics (see for

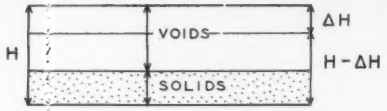


FIG. 1.—Diagram illustrating symbols used in consolidation studies.

example, Tschebotarioff, 1951, p. 105). The symbols illustrated in figure 1 are used.

Settlement is computed by the equation

$$\Delta H = \frac{e_0 - e_f}{1+e_0} H$$

Where

$H$ = the original thickness of a sediment layer
$\Delta H$ = the amount of consolidation or settlement
$e_0$ = the original void ratio
$e_f$ = the final void ratio

It is convenient for the present investigation to substitute into the equation to find $H$, the original thickness,

$$H = \frac{1+e_0}{1+e_f} (H - \Delta H)$$

Where,

$H$-$\Delta H$ = present thickness

In the construction of tables 1–4 typical values were assumed for the properties of the surficial sediments of the deep-sea floor. Using these values as starting points plus pressure-void ratio curves which might be expected in typical clay and calcareous-ooze deposits, the tables were constructed to show the variations of various properties with depth in the sediment (see Hamilton, 1959, for a detailed discussion of the method). By use of the above equation the arbitrary depth increments of tables 1 and 2 can be expanded and added (Cumulative $H$) to show approximately how much original sediment deposition was necessary to obtain any stated present thickness of sediments (tables 3 and 4).

As previously discussed, it appears that a clay under pressure will lithify into a shale at depths between 150 and 700 meters, with a median around 300 meters. Table 1 has been constructed by combining the void ratio-pressure curves for a typical clay (Skempton, 1953; Laughton, 1957) with those of shale (*from* Skeels, 1950) and assum-

TABLE 2.—*Consolidation of a* Globigerina *ooze* (*from Hamilton, 1959*)

| Depth in Sediment (m) | Interval Pressure (Kg/cm²) | Cumul. Pressure (Kg/cm²) | Void Ratio, $e$ | Density sat. (gr/cc) | Submerged wt. (gr/cc) | Porosity, $n$ (%) |
|---|---|---|---|---|---|---|
| 0 | 0 | 0 | 1.90 | 1.62 | 0.57 | 66 |
| 10 | 0.57 | 0.57 | 1.88 | 1.62 | 0.57 | 65 |
| 20 | 0.57 | 1.14 | 1.85 | 1.63 | 0.58 | 65 |
| 30 | 0.58 | 1.72 | 1.82 | 1.63 | 0.58 | 65 |
| 40 | 0.58 | 2.30 | 1.80 | 1.64 | 0.59 | 64 |
| 50 | 0.59 | 2.89 | 1.78 | 1.64 | 0.59 | 64 |
| 100 | 3.00 | 5.89 | 1.68 | 1.66 | 0.61 | 63 |
| 200 | 6.2 | 12.1 | 1.56 | 1.69 | 0.64 | 61 |
| 300 | 6.5 | 18.6 | 1.49 | 1.71 | 0.66 | 60 |
| 400 | 6.7 | 25.3 | 1.44 | 1.72 | 0.67 | 59 |
| 500 | 6.8 | 32.1 | 1.40 | 1.73 | 0.68 | 58 |
| 600 | 6.9 | 38.0 | 1.36 | 1.74 | 0.69 | 58 |
| 700 | 7.0 | 45.0 | 1.32 | 1.76 | 0.71 | 57 |
| 800 | 7.2 | 53.2 | 1.27 | 1.77 | 0.72 | 56 |
| 900 | 7.3 | 60.5 | 1.23 | 1.79 | 0.74 | 55 |
| 1000 | 7.4 | 67.9 | 1.18 | 1.80 | 0.75 | 54 |
| 1500 | 39.0 | 106.9 | 1.01 | 1.87 | 0.82 | 50 |
| 2000 | 42.5 | 149.4 | 0.88 | 1.92 | 0.87 | 47 |

Assumptions: A progressive porosity decrease under pressure with *no* lithification which is probably a rare situation (Hamilton, 1959). Properties: Density of solids: 2.69 gr/cc; water density: 1.05 gr/cc; initial porosity: 66 percent.

Note: This table is derived from Laughton (1957) who ran the test on a *Globigerina* ooze with 54 percent $CaCO_3$.

TABLE 3.—*Thickness of original accumulation necessary to consolidate to present thicknesses in a clay-shale section (from Hamilton, 1959)*

| Depth Interval in Sediment (m) | Present thickness increment (m) | Original accumul., thickness increment (m) | Cumulative thickness of original sed. (m) |
|---|---|---|---|
| 0– 100 | 100 | 155 | 155 |
| 100– 200 | 100 | 190 | 345 |
| 200– 300 | 100 | 205 | 550 |
| 300– 400 | 100 | 220 | 770 |
| 400– 500 | 100 | 230 | 1000 |
| 500– 600 | 100 | 235 | 1235 |
| 600– 700 | 100 | 240 | 1475 |
| 700– 800 | 100 | 245 | 1720 |
| 800– 900 | 100 | 250 | 1970 |
| 900–1000 | 100 | 260 | 2230 |
| 1000–1500 | 500 | 1370 | 3600 |
| 1500–2000 | 500 | 1490 | 5090 |
| 2000–2500 | 500 | 1600 | 6690 |
| 2500–3000 | 500 | 1670 | 8360 |

Note: A transition from clay to shale is assumed to take place at 300 m. There is little significant difference if this transition is not assumed; the consolidation curve for clay shows an original cumulative thickness of 3500 m required to consolidate to a present thickness of 1500 m whereas, with the assumption, 3600 m are required. This table (down to 2000 m) derived from table 1.

ing the transition to take place at a depth of 300 meters. Table 1 and table 3 illustrate the surprisingly large amounts of clay necessary to consolidate and lithify to present thicknesses. A section which is now 500 meters thick would have required approximately 1000 meters of original deposits; a section 2000 meters thick would have required an original deposition of almost 5100 meters.

The tables for calcareous ooze were based on actual tests by Laughton (1957). These tables assume no lithification which is probably a rare condition; it is much more likely that lithification takes place at approximately the same depths as for shale inas-

TABLE 4.—*Thickness of original accumulation necessary to consolidate to present thicknesses in a* Globigerina *ooze section*

| Depth Interval in Sediment (m) | Present thickness increment (m) | Original accumul., thickness increment (m) | Cumulative thickness of original sed. (m) |
|---|---|---|---|
| 0– 100 | 100 | 104 | 104 |
| 100– 200 | 100 | 110 | 214 |
| 200– 300 | 100 | 113 | 327 |
| 300– 400 | 100 | 117 | 444 |
| 400– 500 | 100 | 119 | 563 |
| 500– 600 | 100 | 121 | 684 |
| 600– 700 | 100 | 123 | 807 |
| 700– 800 | 100 | 125 | 932 |
| 800– 900 | 100 | 128 | 1060 |
| 900–1000 | 100 | 130 | 1190 |
| 1000–1500 | 500 | 665 | 1855 |
| 1500–2000 | 500 | 720 | 2575 |
| 2000–3000 | 1000 | 1540 | 4115 |

Note: This table (down to 2000 m) is derived from table 2 wherein the assumption was made that there was a progressive porosity decrease under pressure with *no* lithification; as discussed in the text this is probably a rare situation at deeper depths. After a calcareous ooze has been lithified into a limestone it assumes a strong, rigid structure and it is unlikely that further deposition would cause a marked reduction in porosity.

much as calcareous material is easily cemented by solution and redeposition owing to interstitial water and pressure-chemical effects. As previously noted, calcareous material being a special form of sandy silt or sand-silt-clay responds to pressure less than does clay; consequently there is much less decrease in porosity under gravitational consolidation. A present thickness of 300 meters of calcareous material requires only about 327 meters of original deposits.

## Amounts of Original Deposits at Specific Localities

*Clay-shale sections.*—Challenger Station 26 in the Indian Ocean is on a vast sloping plain south of the Bay of Bengal (Lat. 2°03′ S., Long. 84°58′ E.; Water depth: 2572 fms.) and has a sedimentary thickness greater than the usual deep-sea station (Gaskell, Hill, and Swallow, 1958, p. 60). The first layer is 200 meters thick and overlies 2300 meters of material which has a velocity of 2.47 km/sec at the top; an alternate solution indicates a possible change to 2.7 km/sec at a depth of 1800 meters. These velocities are typical of clay-shale sections (Hamilton, 1959).

The present thickness of 2500 meters under Station 26 required something on the order of 6700 meters of original material to consolidate to the present thickness (table 3). The depositional environment of Station 26 is probably not typical of all deep-sea basins in that the sloping plain may have been in part a turbidity-current depositional feature.

At Challenger Station 44 in the Eastern Atlantic (Lat. 56°34′ N., Long. 21°06′ W.; Water depth: 850 fms.) there are 400 meters of sediment of velocity 1.78 km/sec overlying 1700 meters of material of 2.4 km/sec velocity (at the top of the section); a thin reflecting layer may be present at 400 meters. These thicknesses and velocities indicate a possible clay-shale section. If the present thickness of 1700 meters is a clay-shale section then it took approximately 4120 meters of material to consolidate to the present thickness (table 3).

In a recent paper on seismic refraction results in the Atlantic Ocean, Ewing and Ewing (1959, p. 296) show three stations in and near the Hatteras Abyssal Plain (Stations G-10, G-12, and G-13; located in the area: 25°23′ to 26°25′ N. Lat., 67°26′ to 72°02′ W. Long.; average water depth: 5.47 km). In this area there is an average of 320 meters of sediment (velocity 1.70 km/sec) overlying second layers with an average velocity of 2.16 km/sec (at the top of the layer) and an average thickness of 883 m (fig. 2). These velocities and thicknesses indicate a probable clay-shale section. If this is a clay-shale section, it required an average thickness of approximately 2780 meters of material to consolidate to the present average thickness of 1203 meters (table 3).

In the northwest Atlantic abyssal plain, Ewing and Ewing (1959. p. 296; Sta. A-157-29 at about 37°10′ N. Lat. 65°03′ W. Long.; water depth 5.05 km) record a first layer thickness of 440 meters (velocity 1.62 km/sec) overlying a second layer of 1090 meters thickness with a velocity of 1.82 km/sec at the top of the section. If this is a clay-shale section (as is probable) then it took approximately 3690 meters of original deposits to consolidate to the present thickness of 1530 meters (table 3).

In the Norwegian Basin. Ewing and Ewing (1959, p. 296) record two stations (F-7 and F-9) where an average thickness of 285 meters of sediment (velocity, 1.75 km/sec) overlies a second layer averaging 695 meters thick with an average velocity (at the top) of 1.94 km/sec. If this is a clay-shale section, it would have required approximately 2180 meters of original deposits to consolidate to the present thickness of 980 meters (table 3).

*Calcareous ooze-limestone sections.*—In the east equatorial Pacific, Capricorn Stations 18, 19. 20, 21, 22, and 23 (Raitt, 1956) are all well into the calcareous ooze area away from ridges and boundaries of present-day red clay deposition. The average thickness of the first layer is 260 meters and the average thickness of the second layer is 1090 meters, a total thickness of 1350 m; the velocity in the second layer averages 5.09 km/sec. If it is assumed that calcareous ooze deposition began in this area 100 million years ago during the Late Cretaceous (when planktonic Foraminifera evolved) and that the eastern Pacific Ocean Basin is older than Late Mesozoic, then it is very probable that the lower part of the sediments is a clay

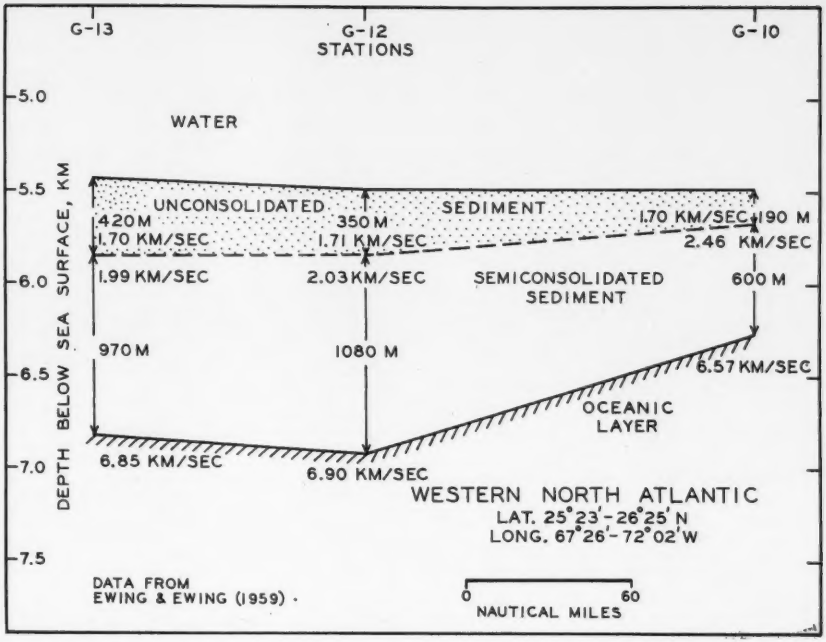


FIG. 2.—Seismic refraction section across the south end of the Hatteras Abyssal Plain in the Western North Atlantic (from Ewing and Ewing, 1959).

shale (Hamilton, 1959). An average rate of deposition of calcareous material of 1.0 cm/1000 years for 100 million years would result in an original thickness of 1000 meters of calcareous ooze.

Tables 2 and 4 illustrate the consolidation of a calcareous ooze without lithification. As discussed in a previous paper (Hamilton, 1959), limestone does not markedly decrease in porosity under pressure as does shale; numerous sections from oil wells show that this is essentially true. In the east equatorial Pacific at the above cited Capricorn Stations the average thickness of the first layer is 260 meters. If it is assumed that the calcareous material is lithified into limestone at this depth and that the material below does not decrease in porosity under pressure, then it is apparent that one should not use the tables for unlithified material to determine the present thickness of any stated amount. It should be more accurate to consider that there is a topmost layer which consolidates along expectable curves and, when lithified, does not further decrease in porosity. In this case, a present thickness of 260 meters would be derived from an original accumulation of 280 meters; 1000 meters then would decrease to 930 meters.

If the present thickness of sediment in the east equatorial Pacific, at the stations cited above, is 1350 meters and there is a lower shale section, then the whole section should be divided as follows: the original 1000 meters of calcareous material has consolidated and lithified into 930 meters of calcareous ooze and limestone (260 meters of ooze and 670 meters of limestone) and there is an underlying shale section of 420 meters. Simple trigonometry and Snell's Law indicate no significant additional thickness of the whole section because of assuming a third layer, shale, rather than two cal-

TABLE 5.—*Estimates of the amounts of solids in original deposits at selected localities*

| Area | Station | Sediment Thickness (m) | | Amount (thickness) of solids (m) |
|---|---|---|---|---|
| | | Present | Original | |
| East Pacific (ave) | MidPac 18, 19, 20, 21, 22, 23 | 1350 | 2090 | 630 |
| West Atlantic | G-10, 12, 13 | 1203 | 2780 | 835 |
| West Atlantic | A-157-29 | 1530 | 3690 | 1110 |
| Indian Ocean | Challenger 26 | 2500 | 6700 | 2010 |
| East Atlantic | Challenger 44 | 1700 | 4120 | 1240 |
| Norwegian Sea | F-7, 9 | 980 | 2180 | 655 |

careous layers (the second layer being the lithified equivalent of the first layer).

The measured velocity in the second layer of the cited Capricorn Stations averages 5.09 km/sec. If the second layer is limestone, this velocity implies a porosity of approximately 8 percent (Wyllie, Gregory, and Gardiner, 1956); if so, then the intergranular pressure at the base of the limestone is approximately 117 kg/cm$^2$.

Given a shale section of 420 meters under a pressure at the top of 117 kg/cm$^2$, one can enter the pressure-void ratio curves and the depth-pressure curves (Hamilton, 1959) and determine the average void ratio of the section. Assuming the initial properties of the clay to be the same as shown in table 1, it is then possible to compute that the present thickness of 420 meters was derived from an original clay deposition of 1090 meters, which, when added to the previously estimated 1000 meters of calcareous material, comprises a total section of 2090 meters of original deposits.

## THE AMOUNT OF SOLIDS IN DEEP-SEA SEDIMENTS

Geochemical and other considerations (Kuenen, 1950, 386–398; Revelle, 1954, p. 327) indicate that the thickness of sediments on the deep-sea floor should be of the order of 1 to 3 km and that the amount of solids should be 2½ km for the higher value (Kuenen, 1950, p. 398). These are rough, averaged calculations and do not imply that there should be this thickness of solids everywhere. The hypotheses, calculations, and measurements previously discussed allow some generalized calculations of the thickness of solid material at any one station. Making a general assumption that the average porosity of all of the original deposition was 70 percent (30 percent solids) the amounts of solids shown in table 5 would be present at the stations discussed above.

It can be seen that even with a conservative estimate for the amount of solids in the original sediments (30 percent), the total amounts of solids for these sections fall close to the estimated thicknesses of solids required by the geochemical balance (1–3 km).

## AGES OF OCEAN BASINS

### Rates of Deposition

Obviously, we have here an interesting method of estimating the ages of ocean basins if only valid rates of deposition were available.

Broad generalizations about the rates of sedimentation on the deep-sea floor have little validity if single generalized figures are given for the deposition of red clay and calcareous ooze. Schott's statement of caution (1939, p. 414) in generalizing from his values for sediments in the equatorial Atlantic and Indian Oceans is amply borne out by later investigations. Ericson and others (1956), in discussing the rates of deposition on banks, ridges, rises, slopes, and in basins of the Atlantic, find values for the rates of sediment deposition varying from 2.2 to 270 cm/1000 years; in general the rates are low on highs and high in basins which can receive turbidity current deposits and sediments swept from highs. A sharp distinction must be made between the rate of deposition and the rate of accumulation (as was done by Arrhenius, 1952). In the Pacific, for example a flat-topped seamount in the Mid-Pacific Mountains has about an inch of recent calcareous sediment over pure Eocene calcar-

eous ooze with no observable discontinuity; the thickness of sediment here has little to do with the rate of deposition (Hamilton 1953). Highs of the deep South Pacific sea floor have ancient sediments at the surface, and calcareous ooze collects in thick deposits under the current divergences of the equatorial Pacific (Riedel, 1957; Arrhenius, 1952).

Very few estimates of the rate of deposition of deep-sea sediments have been made which can be used to compute pre-Pleistocene sediment thicknesses. Volchok and Kulp (1957) thought that the post-Pleistocene rate of deposition is about three times the rate of earlier deposition. Goldberg and Arrhenius (1958) thought that Recent rates are significantly greater than Tertiary rates and might be on the order of 3 or 4 times greater. Piggot and Urry (1942) also thought that the Recent rate was more than that for the past 300,000 years (×4 in the case of a Caribbean core). At the present time it appears that post-Pleistocene values should be divided by at least three to estimate pre-Recent (or pre-Pleistocene values).

Both Shepard (1948, p. 306) and Kuenen (1950, p. 383) allow 40 percent for core shortening when correcting Schott's (1939) and other gravity-core values of sediment thickness. Schott's average calcareous sediment thickness was 24 cm for the Equatorial Atlantic, and 11.7 cm for the Indian Ocean. Careful experiments of core recovery using different sizes of core tubes and different methods of forcing the sampler into the sediment have shown that the upper parts (0 to 15 to 25 in) of gravity cores should be relatively undisturbed and that *no* allowance should be made for core shortening for the topmost part of such cores (Hvorslev, 1949, p. 83–119). Schott's (1939), Bramlette and Bradleys' (1940) and other estimates based on a thickness of less than 15 to 25 in of sediment from the tops of gravity cores should not be corrected for core shortening.

Schott allowed 20,000 years for the time from the end of the Pleistocene to the present; Bramlette and Bradley allowed 9000 years. Ericson and others (1956) have shown that over most of the North Atlantic and adjacent seas the climatic change occurred approximately 11,000 years ago; this more recent estimate should be used to correct earlier estimates of rates of deposition which used the stratigraphic method (that is, a thickness of sediment deposited in so-many years since the end of the Pleistocene).

Table 6 summarizes some values for pre-Recent rates of deposition based on the following assumptions: (1) the rate for post-Pleistocene deposition is three times that before the Ice Ages, (2) no core shortening should be allowed in the uppermost parts of gravity cores, and (3) the last great climate change at the end of the last ice age occurred 11,000 years ago. Some of the radioactive data are for the pre-Recent and need no such corrections.

TABLE 6.—*Pre-Recent rates of deposition*

| Area and reference | Rate reported (cm/1000 yrs) | Pre-Recent corrected rate (cm/1000 yrs) |
|---|---|---|
| Calcareous Ooze | | |
| Equatorial Atlantic (Schott, 1939) | 1.2 | 0.71 |
| Indian Ocean (Schott, 1939) | 0.59 | 0.36 |
| North Atlantic (Bramlette and Bradley, 1940) | 3.8 | 1.04 |
| Equatorial Pacific (Arrhenius, 1952) | 3.2 | 1.07 |
| Carribbean (Piggot and Urry, 1942) | 4.0 | 1.33 |
| Carribbean (Volchok and Kulp, 1957) | 0.4–0.8 | 0.4–0.8 |
| Red Clay | | |
| Eastern Pacific (Goldberg and Koide, 1958) | 0.1–0.2 | 0.05 |
| Equatorial Atlantic (Schott, 1939) | 0.86 | 0.52 |
| Eastern Pacific (Piggot and Urry, 1942) | 0.5 | 0.5 |
| South Pacific (Volchok and Kulp, 1957) | 0.10–0.20 | 0.1–0.2 |
| South Pacific (Goel and others, 1957) | 0.55 | 0.55 |
| Equatorial Pacific (Arrhenius, 1952) | 0.29 | 0.1 |

It is apparent that the rate of deposition varies from place to place (as it should) and that the magnitude of deposition in the deep sea is of the order of 0.4 to 1.3 cm/1000 years for calcareous ooze and 0.05 to 0.5 cm/1000 years for red clay for the pre-Recent IF the present rates and determinations can be extrapolated into earlier geologic time. These ranges might be used as a first approximation in estimating time in thickness computations.

## Age Computations

As previously discussed, the rates of deposition vary widely from area to area and from the Recent to Pleistocene time; consequently it is to be expected that these rates will vary through geologic time. The total thickness of sediments at any one locality will be dependent on the age of the ocean basin, the rates of deposition and accumulation, and geologic history of the locality (for example, whether or not there has been any activity leading to redeposition of sediment because of slumping or lack of accumulation because of removal by currents). For these reasons it is a sterile exercise to average thicknesses of sediments over large areas and from diverse environments. It is necessary to pick a single station or stations representing, as far as possible, a single environment or area. With this in mind it is interesting to compute the ages of ocean basins using the information at hand.

The specific localities previously discussed will serve to illustrate age computations based on the amounts of original sediments and assumed average rates of deposition.

At Challenger Station 26 in the Indian Ocean it is estimated that there are at present 2500 meters of material which is probably clay-shale. This present thickness required 6700 meters of original sediment thickness. Sediment rates on this sloping plain south of the Bay of Bengal are apt to be higher than for deep-sea clay. If the material was deposited at an average rate of 1 cm/1000 years, the basin in this area is approximately 670 million years old; if the material was deposited at a rate of 2 cm/1000 years, the basin is 335 million years old.

At Challenger Station 44 in the Eastern Atlantic the present thickness of sediments is considered to be 1700 meters. If this material is a clay-shale section, as appears likely, it took 4120 meters of original sediment thickness to consolidate to the present thickness. If the clay was deposited at a rate of 1 cm/1000 years, the basin in this area is approximately 412 million years old; if the clay was deposited at a rate of 0.5 cm/1000 years, the basin is 824 million years old.

Capricorn Stations 18, 19, 20, 21, 22, and 23 in the eastern equatorial Pacific are all from an environment of calcareous ooze deposition. As discussed in the previous section, there is a possibility that there are approximately 1000 meters of original calcareous deposits overlying a shale section which would have required approximately 1090 meters of original clay deposits. If the calcareous material was deposited in 100 million years (at a rate of 1 cm/1000 years), the age of the ocean basin is figured at 318 million years if the lower layer of clay was deposited at a rate of 0.5 cm/1000 years, at 536 million years if the rate was 0.25 cm/1000 years, and 1190 million years if the rate was 0.1 cm/1000 years.

In the Western Atlantic Basin in the south end of the Hatteras Abyssal Plain, the three stations previously discussed indicate an average first and second layer thickness of 1203 meters which, if a clay-shale section, would have required an average thickness of approximately 2780 meters of original sediments. The material comprising this section is undoubtedly clay and calcareous ooze and turbidity-current deposits, but if these 2780 meters of sediments were deposited at an average rate of 1 cm/1000 years the basin is 278 million years old, if at an average rate of 0.5 cm/1000 years the basin is 556 million years old, or if at a rate of 0.1 cm/1000 years the basin is 2780 million years old.

Obviously the above age computations are nebulous in that the ancient rates of sedimentation are at best "guesstimates." The results of this method, however, fall into the category of evidence which indicates ancient age of the ocean basins (Paleozoic to Pre-Paleozoic).

The methods and computations of this paper indicate the possible power of the application of soil mechanics techniques to some of the problems of deep-sea sedimentation. When it has been decided whether or

not the second layers of seismic literature are, in fact, the lithified equivalents of the material now being deposited, it will be possible to apply these methods to determine the total volume of sediments, estimate the amounts of solids in these sediments, and (with age-dated borings from a few localities to give overall sedimentation rates) estimate the ages of the ocean basins.

Acknowledgments

The writer appreciates critical reading and helpful comments from G. O. S. Arrhenius, E. C. Buffington, G. H. Curl, R. S. Dietz, R. L. Fisher, E. D. Goldberg, J. B. Hersey, H. W. Menard, D. G. Moore, R. W. Raitt, R. Revelle, C. J. Shipek, G. G. Shor, Jr., and G. Shumway.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 380–403
FIGS. 1–10, SEPTEMBER, 1960

# TEXTURE AND MINERALOGY OF THE RECENT SANDS OF THE GULF COAST[1]

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## ABSTRACT

The objectives of this investigation are to determine the major transportation paths for the Recent beach sands of the Gulf of Mexico and to relate the textures and mineralogy of the Recent sands to their depositional environments.

Approximately two hundred Recent sand samples were collected from the rivers and beaches of the Gulf Coast states. Mechanical analyses were made to determine the grain size and sorting. Spectrochemical methods were used to determine chemical composition, and thin-section Rosiwal analyses were used to determine mineralogical composition.

The major conclusions of this investigation are as follows:

1.—The Recent beach sands east of the Mississippi River Delta are mature quartz sands which contain practically no feldspar and a mature heavy-mineral suite rich in staurolite and kyanite. The Recent beach sands west of the Mississippi River are feldspathic. The feldspar abundance of these sands decreases from the Mississippi Delta area westward in the direction of prevailing longshore currents, mainly by dilution. The heavy-mineral suite of the Recent beach sands from Louisiana and east Texas contains abundant unstable minerals; among the resistant heavy minerals, garnet predominates over staurolite and kyanite.

2.—The relative abundance of unstable heavy minerals which can easily be decomposed during weathering or by postdepositional changes is not a reliable criterion for paleogeographic interpretations. On the other hand, feldspar abundance and ratio of garnet to staurolite plus kyanite, can be used to recognize ancient petrologic provinces.

## INTRODUCTION

Petrologic provinces have been recognized in the Recent sediments of the northern Gulf Coast on the basis of heavy-mineral investigations (Goldstein, 1942; Bullard, 1942). Casual observations have shown that the sands from different petrologic provinces are also different in bulk composition (R. H. Nanz, personal communication). The beach sands east of the Mississippi River Delta are orthoquartzitic and are characterized by a heavy-mineral suite rich in staurolite and kyanite. The beach sands west of the Mississippi River Delta are mostly feldspathic and are characterized by a heavy-mineral suite rich in amphiboles; these sands are similar, mineralogically, to the Mississippi River sand (Russell, 1937, and Goldstein, 1942). This similarity led to the conclusion that the Mississippi River sediment swept westward by longshore currents provided much of the source material for the beach sands west of the Mississippi River Delta (Goldstein, 1942). However, mineralogic differences between the Mississippi River sand and the beach sands have also been noted; the beach sands, for example, contain less feldspar and a more mature heavy-mineral suite than the Mississippi River sand (Goldstein, 1942; Nanz, unpublished data). The question was therefore raised whether such differences in mineralogy can be attributed to abrasional loss of "less resistant" minerals during the westward longshore transport or to dilution due to contributions of other source materials.

Investigations of Recent river sands (for example, Russell, 1937; Van Andel, 1950) have indicated that selective destruction of the "less resistant" minerals appears to be minor during river transport. Whether abrasional loss during longshore transport of beach sediments is significant remains in doubt.

The present investigation was initiated in 1954 in order to evaluate whether the mineralogic differences between the Mississippi River sand and the beach sands of the western Gulf Coast have resulted primarily from abrasion or from dilution. It is further hoped that the mineralogy and textures of

[1] Publication no. 218, Shell Development Company. Manuscript received October 12, 1959.

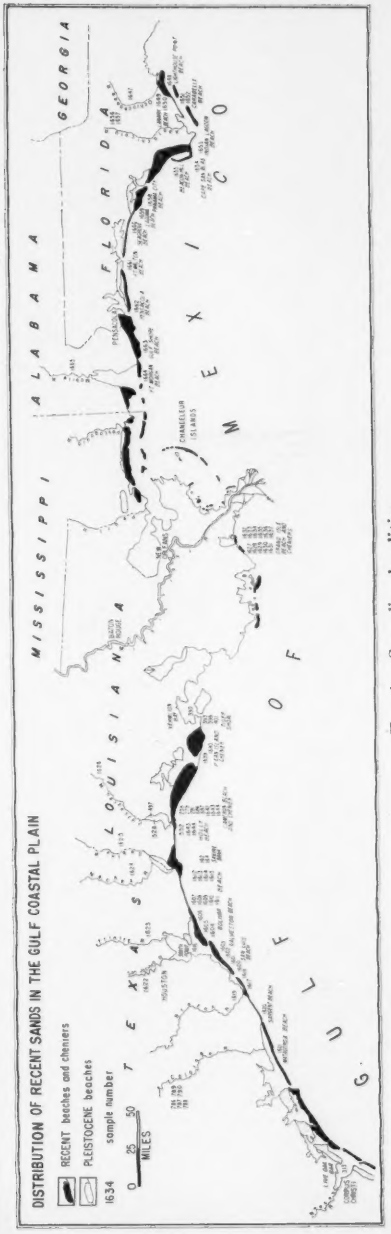


FIG. 1.—Sampling localities.

the Gulf Coast Recent sands can be related to their depositional environments. A correct understanding of the origin of the Recent sands should form a basis for successful interpretation of the genesis of many ancient sandstones.

Mechanical analyses were used to determine grain size and sorting coefficient of the sand samples. The chemical composition of the sands was determined by spectrographic methods, and thin-section Rosiwal analyses were used to determine the feldspar abundances. Practically all heavy-mineral-analysis data were taken from the literature and were integrated here to complement the results of the present investigation.

### SAMPLING AND TREATMENT OF SAMPLES

Six sampling trips were taken from March to August 1954. Grab samples were taken from the beaches and river point bars of Texas, Louisiana, Alabama, and western Florida. Samples of offshore marine sands were furnished by R. J. LeBlanc of Shell Development Company. The sampling localities are shown in figure 1. The treatment of samples is outlined in figure 2.

### SIZE AND SORTING OF RECENT SANDS

#### Recent Beach Sands of the Texas Coast

Two sampling trips were made along the Texas coast. Sampling localities on the beaches were first spaced at 5-mile intervals and then at 1-mile intervals. Table 1 shows the median size ($d_m$ in mm) and sort-coefficient (So) of the sands sampled on the first trip. Results of the second trip, which are too bulky to be presented here, confirm these values. The median diameters of the sands range from 0.089 to 0.221 mm. These sands are thus all fine and very fine sands, according to Wentworth's (1922) grade scale. The sands of the Sargent and Matagorda beaches are, in general, coarser grained than the beach sands of Bolivar Peninsula and Galveston Island. Also, invariably the beach sands are finer grained near the mouth of a river (fig. 3). These beach sands are all extremely well sorted. Variations in sorting along the strike of the coast are minor and are not systematic. The sorting coefficients (So) of the sands range from 1.07 to 1.26.

Sands from a Recent abandoned beach ridge, or chenier, near Sabine Pass and from

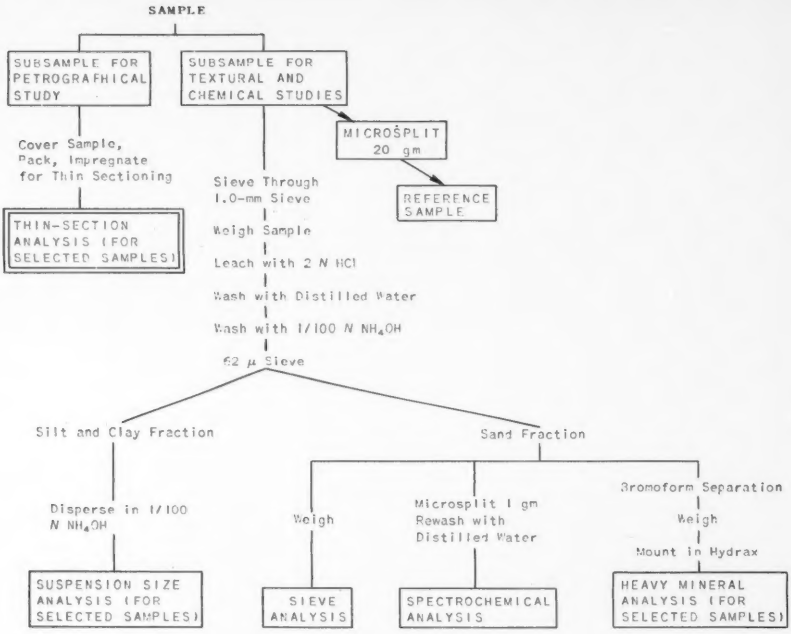


FIG. 2.—Treatment of samples.

a Pleistocene barrier beach near Smith Point are similar in grain size and sorting to the Recent beach sands. The chenier sand has a median diameter of 0.140 mm and an So of 1.14; the Pleistocene sand has a median diameter of 0.150 mm and an So of 1.14.

TABLE 1.—*Grain size and sorting of Recent beach sands of Texas*

| Sample Number | Location | $d_m$ (in mm) | So |
|---|---|---|---|
| 1600 | Galveston Island | 0.135 | 1.16 |
| 1601 | Galveston Island | 0.130 | 1.16 |
| 1602 | Galveston Island | 0.140 | 1.16 |
| 1603 | Galveston Island | 0.131 | 1.14 |
| 1604 | Bolivar Beach | 0.148 | 1.16 |
| 1605 | Bolivar Beach | 0.157 | 1.18 |
| 1606 | Bolivar Beach | 0.170 | 1.19 |
| 1607 | Bolivar Beach | 0.149 | 1.13 |
| 1608 | Bolivar Beach | 0.150 | 1.13 |
| 1609 | Bolivar Beach | 0.134 | 1.13 |
| 1610 | Bolivar Beach | 0.145 | 1.11 |
| 1611 | Bolivar Beach | 0.145 | 1.12 |
| 1612 | Bolivar Beach | 0.105 | 1.15 |
| 1617 | San Luis Beach | 0.125 | 1.13 |
| 1618 | San Luis Beach | 0.159 | 1.19 |
| 1620 | Sargent Beach | 0.229 | 1.15 |
| 1621 | Matagorda Beach | 0.168 | 1.13 |

## Recent Beach Sands of the Louisiana Coast

Sand beaches of the Louisiana coast are of limited lateral extent; they are separated by coastal mud flats and marshlands. Selected sand samples were taken from present-day beaches and from Recent abandoned beach ridges. The median diameters and sorting coefficients of these samples are shown in table 2.

The size and degree of sorting of the Louisiana beach sands are of approximately the same order of magnitude as those of the Texas beach sands.

The median diameters of the sands from cheniers (abandoned beach ridges) are comparable to those of the sands on beaches today, except for sample 1638 which was collected from Front Ridge of Pecan Island.

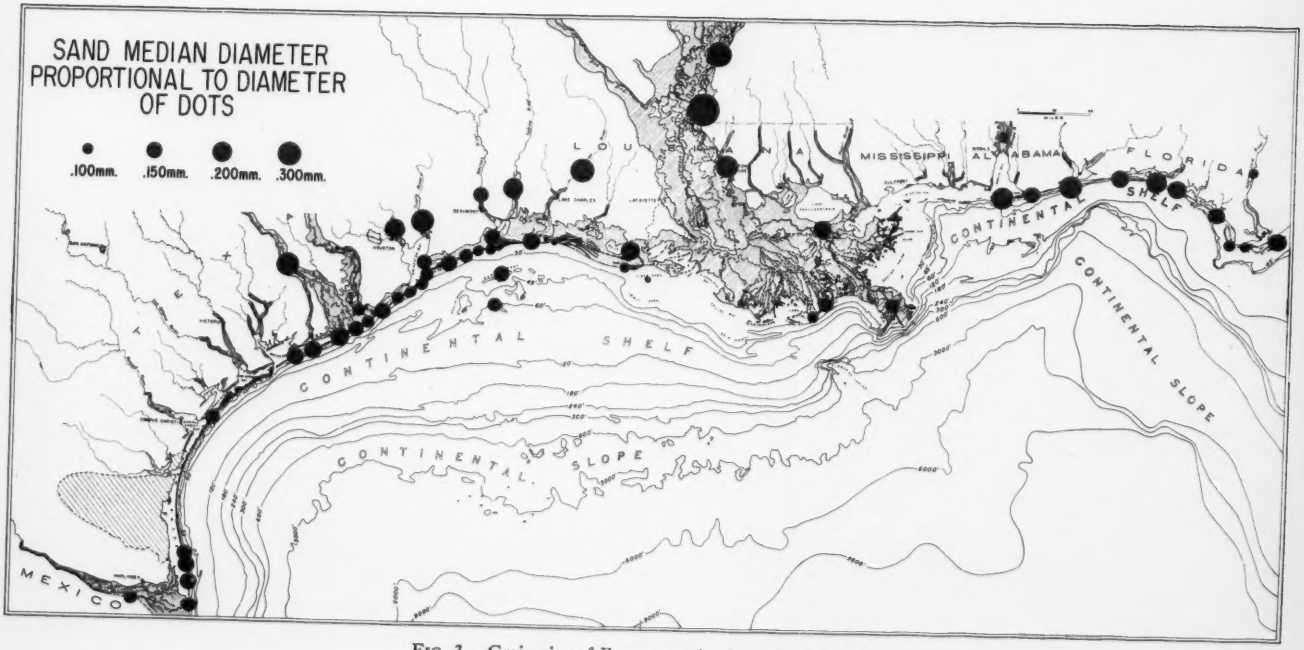


FIG. 3.—Grain size of Recent sands of the Gulf Coast.

TABLE 2.—*Grain size and sorting of Recent beach sands of Louisiana*

| Sample Number | Location | $d_m$ (in mm) | So |
|---|---|---|---|
| 1630 | Grand Isle | 0.153 | 1.09 |
| 1631 | Grand Isle | 0.151 | 1.09 |
| 1632 | Grand Isle | 0.151 | 1.09 |
| 1633 | Grand Isle | 0.150 | 1.11 |
| 1634 | Grand Isle | 0.150 | 1.13 |
| 1635 | Grand Isle | 0.135 | 1.15 |
| 1636 | Grand Isle | 0.151 | 1.06 |
| 1637 | Grand Isle | 0.149 | 1.09 |
| 1642 | Cameron Beach | 0.090 | 1.14 |
| 1644A | Cameron Beach | 0.090 | 1.12 |
| 1644B | Cameron Beach | 0.087 | 1.13 |
| 1645 | Holly Beach | 0.185 | 1.24 |
| 1646 | Holly Beach | 0.195 | 1.24 |

The chenier sands are apparently less well sorted (higher So) than the modern beach sands. However, Nanz (1954) pointed out that interstitial material (tail fraction) in ancient sands may have been introduced subsequent to deposition and that the $d_m$ and So of the framework fraction should be used to represent the size and sorting of ancient sands. As anticipated, the sorting coefficients of the framework fraction of the chenier sands are comparable to those of the beach sands of today.

## Recent Beach Sands of the Eastern Gulf Coast

Sand beaches extend almost continuously from Apalachicola Bay, Florida to Mobile Bay, Alabama. The beach sands are fine to medium. Those immediately west of the Apalachicola River are finer grained than the sands of interdeltaic beaches, which again confirms the observations noted along the Texas coast that the median diameter of a beach sand is finer near the mouth of a river (fig. 3).

TABLE 3.—*Grain size and sorting of Recent beach sands of the Eastern Gulf*

| Sample Number | Location | $d_m$ (in mm) | So |
|---|---|---|---|
| 1648 | Light House Point Beach | 0.267 | 1.26 |
| 1650 | Lanark Beach | 0.310 | 1.39 |
| 1651 | Carabelle Beach | 0.235 | 1.27 |
| 1653 | Indian Lagoon | 0.145 | 1.14 |
| 1654 | Cape San Blas Beach | 0.176 | 1.09 |
| 1655 | Beacon Hill Beach | 0.239 | 1.16 |
| 1658 | Panama City Beach | 0.245 | 1.15 |
| 1659 | Laguna Beach | 0.265 | 1.14 |
| 1660 | Seagrove Beach | 0.245 | 1.15 |
| 1661 | Ft. Walton Beach | 0.240 | 1.13 |
| 1662 | Pensacola Beach | 0.263 | 1.16 |
| 1663 | Gulf Shore Beach | 0.233 | 1.17 |
| 1664 | Ft. Morgan Beach | 0.252 | 1.20 |

The beach sands west of the Apalachicola River are all extremely well sorted; the sorting coefficients range from 1.09 to 1.20, comparable to those of the Texas beach sands. The beach sands east of the Apalachicola River are less well sorted, however, with sorting coefficients ranging from 1.26 to 1.39 (table 3).

## Recent River Sands

Representative samples of point-bar sands were collected from the Brazos, San Jacinto, Trinity, and Neches Rivers in Texas, the Sabine and Calcasieu Rivers in Louisiana, the Mobile River in Alabama, and the Apalachicola River in Florida. The median diameters of these point-bar sands range from 0.111 to 0.325 mm, and sorting coef-

TABLE 4.—*Grain size and sorting of Recent river sands of the Gulf Coast*

| Sample Number | Location | $d_m$ (in mm) | So |
|---|---|---|---|
| 1619C | Brazos River | 0.120 | 1.23 |
| 1622 | San Jacinto River | 0.210 | 1.31 |
| 1623 | Trinity River | 0.250 | 1.26 |
| 1624 | Neches River | 0.140 | 1.32 |
| 1625 | Sabine River | 0.260 | 1.50 |
| 1626 | Calcasieu River | 0.325 | 1.20 |
| 1656A | Apalachicola River | 0.126 | 1.24 |
| 1656B | Apalachicola River | 0.111 | 1.24 |
| 1656C | Apalachicola River | 0.125 | 1.31 |
| 1657 | Apalachicola River | 0.129 | 1.16 |
| 1665A | Mobile River | 0.142 | 1.75 |
| 1665B | Mobile River | 0.152 | 1.25 |
| 1788A | San Bernard River Delta | 0.074 | 1.39 |
| 7788B | San Bernard River Delta | 0.082 | 1.33 |
| 1788C | San Bernard River Delta | 0.105 | 1.34 |
| 1788D | San Bernard River Delta | 0.093 | 1.40 |
| 1788E | San Bernard River Delta | 0.112 | 1.28 |
| 1788F | San Bernard River Delta | 0.109 | 1.21 |
| 1788G | San Bernard River Delta | 0.084 | 1.73 |

TABLE 5.—*Size and sorting of Mississippi River sands**

| Location (Miles below Cairo) | No. of Samples | $d_m$ (mm) | | So | |
|---|---|---|---|---|---|
| | | Range | Average | Range | Average |
| 0– 100 | 10 | 0.263–1.616 | 0.868 | 1.22–2.54 | 1.54 |
| 100– 200 | 10 | 0.318–1.019 | 0.670 | 1.18–2.00 | 1.50 |
| 200– 300 | 10 | 0.302–0.754 | 0.510 | 1.18–1.87 | 1.35 |
| 300– 400 | 10 | 0.205–0.885 | 0.480 | 1.12–2.79 | 1.45 |
| 400– 500 | 10 | 0.239–1.541 | 0.580 | 1.10–2.21 | 1.32 |
| 500– 600 | 10 | 0.241–0.638 | 0.391 | 1.16–1.51 | 1.25 |
| 600– 700 | 8 | 0.090–0.591 | 0.322 | 1.16–1.44 | 1.23 |
| 700– 800 | 8 | 0.173–0.562 | 0.349 | 1.14–1.30 | 1.20 |
| 800– 900 | 10 | 0.199–0.359 | 0.272 | 1.10–1.14 | 1.12 |
| 900–1000 | 9 | 0.174–0.328 | 0.224 | 1.15–1.21 | 1.28 |
| 1000–Gulf | 8 | 0.115–0.202 | 0.157 | 1.14–1.34 | 1.18 |

* Averages, computed from the data published by the U. S. Corps of Engineers (1935).

ficients range from 1.16 to 1.75 (table 4). In general, the point-bar sands collected nearer to the mouths of the rivers are finer grained (fig. 3). The sands from the San Bernard River Delta are the finest.

Mississippi River sand samples from Cairo, Illinois, downstream to the Gulf of Mexico were collected and analyzed by the U. S. Corps of Engineers, Waterways Experiment Station (1935) at Vicksburg, Mississippi. From those analyses, median diameters ($d_m$) and sorting coefficients (So) for approximately 100 samples spaced at about 10-mile intervals were computed by the writer. Average values of $d_m$ and So for 100-mile intervals, calculated from these results, are listed in table 5. The Mississippi River sands within 500 miles downstream from Cairo are medium to coarse grained and are "poorly sorted" as compared with the Recent beach sands of the Gulf region. The River sands become finer grained and better sorted downstream. In the last 100 miles, the River sands have median diameters and sorting coefficients which are very similar to those of the beach sands on Grand Isle.

## MINERALOGICAL COMPOSITION OF RECENT SANDS

### Mineralogy of Recent Beach Sands

*Heavy-mineral provinces in Recent sands recognized on the basis of differences in total heavy-mineral composition of the sands.*—Four petrologic provinces exist along the northern and western shores of the Gulf of Mexico:

1.—Eastern Gulf province.
2.—Mississippi River province.
3.—Western Gulf province.
4.—Rio Grande province.

These petrologic provinces were first recognized on the basis of differences in total heavy-mineral composition (Goldstein, 1942 Bullard, 1942). The eastern Gulf province is characterized by a relatively high content of ilmenite, staurolite, kyanite, zircon, tourmaline, and sillimanite, and by low percentages of magnetite, amphiboles, and pyroxenes. The Mississippi River province is characterized by abundant amphiboles, dolomite, pyroxenes, epidote, ilmenite, and biotite. The western Gulf province has a heavy-mineral assemblage similar to that of the Mississippi River province, but pyroxenes and ilmenite are less common and leucoxene is more abundant. The Rio Grande province has an assemblage of heavy minerals characterized by high percentages of basaltic hornblende and pyroxene.

*Heavy-mineral provinces in Recent sands recognized on the basis of differences in S/U ratio of the sands.*—Differences in heavy-mineral compositions of the Recent sands along the northern shores of the Gulf of Mexico are directly related to the maturity of these sediments. The ratio of stable (resistant to weathering and/or to diagenetic changes) to unstable (nonresistant to such changes) nonopaque heavy minerals is a convenient measure of the maturity of Recent sands (Bullard, 1942). This ratio will hereafter be termed the *S/U* ratio. Pyrox-

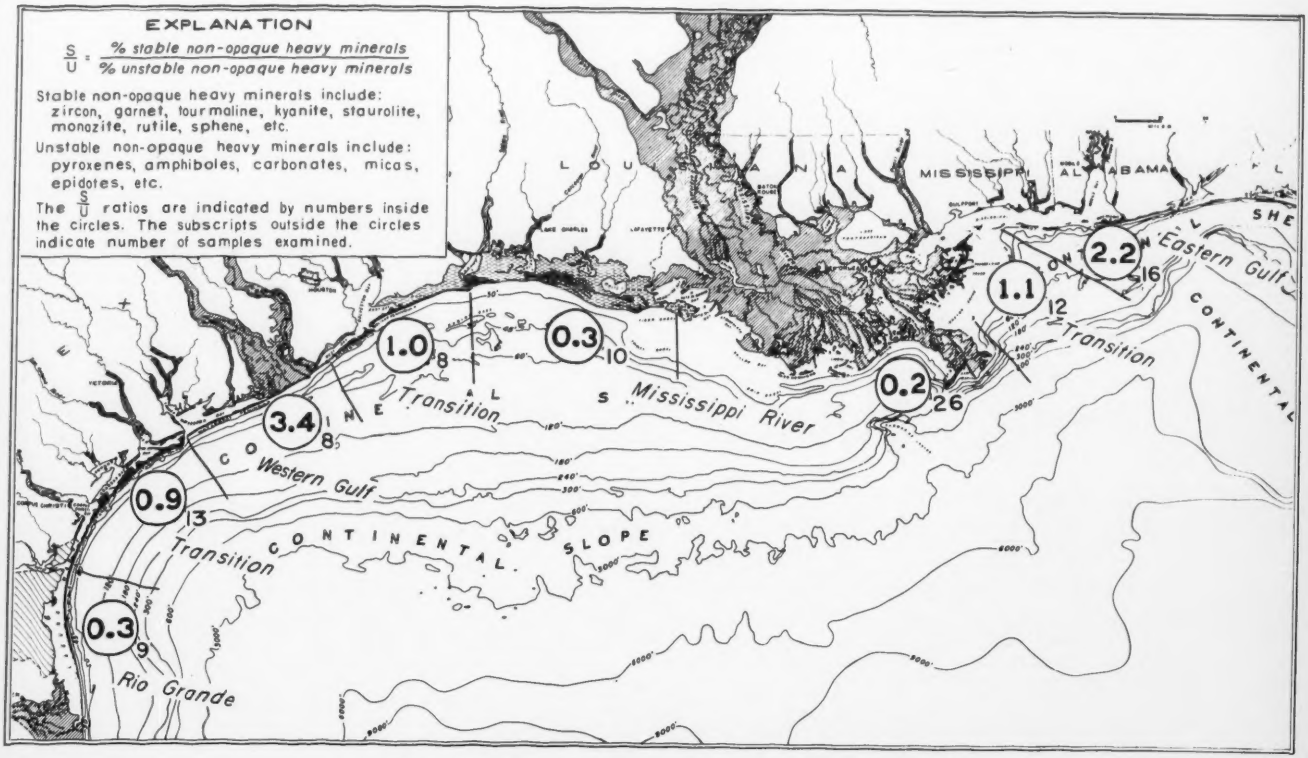
EXPLANATION
S/U = % stable non-opaque heavy minerals / % unstable non-opaque heavy minerals
Stable non-opaque heavy minerals include: zircon, garnet, tourmaline, kyanite, staurolite, monazite, rutile, sphene, etc.
Unstable non-opaque heavy minerals include: pyroxenes, amphiboles, carbonates, micas, epidotes, etc.
The S/U ratios are indicated by numbers inside the circles. The subscripts outside the circles indicate number of samples examined.
LOUISIANA
MISSISSIPPI
ALABAMA
1.0 8
0.3 10
3.4 8
0.9 13
0.3 9
0.2 26
1.1 12
2.2 16
Transition
Mississippi River
Western Gulf
Eastern Gulf
CONTINENTAL SLOPE
Rio Grande

FIG. 4.—Average *S/V* ratios in the sands of the Gulf Coast.

enes, amphiboles, dolomite and epidote have generally been considered as relatively unstable minerals, and garnet, staurolite, kyanite, tourmaline, and zircon have been considered as relatively stable minerals (Pettijohn, 1949, p. 489). The $S/U$ ratios for the Gulf Recent sands, computed from the data of Goldstein (1942) and Bullard (1942), are shown in figure 4.

The petrologic provinces recognized on the basis of total heavy-mineral compositions have significantly different $S/U$ ratios (fig. 4). Sands of the Mississippi River and Rio Grande provinces are comparatively immature, with an $S/U$ ratio of 0.2 to 0.3. Sands of the eastern and western Gulf provinces are mature, with an $S/U$ ratio of 2 to 3. Sands in the transition zones are intermediate in maturity, with an $S/U$ ratio of about 1.

*Heavy-mineral provinces in Recent sands recognized on the basis of differences in $G/(S+K)$ ratios of the sands.*—The five most common stable nonopaque heavy minerals in the Recent sands of the Gulf Coast region are zircon, tourmaline, garnet, staurolite, and kyanite. Of these five, the percentages of garnet and of staurolite plus kyanite vary significantly in different parts of the Gulf Coast. Whereas the eastern Gulf is extremely low in garnet and very rich in staurolite and kyanite, the petrologic provinces west of the Mississippi River Delta all have a heavy-mineral suite richer in garnet and poorer in staurolite and kyanite. The garnet/(staurolite+kyanite) ratio, hereafter termed the $G/(S+K)$ ratio, has been computed from the data of Goldstein (1942), Bullard (1942), and Richardson (1948); the results are shown in figure 5. This ratio is less than 0.1 in the Recent sands of the eastern Gulf province but is greater than 1 in the petrologic provinces west of the Mississippi River Delta.

*Petrologic provinces recognized on the basis of feldspar abundance.*—The differences in maturity of the Gulf Coast Recent sands, as revealed by heavy-mineral investigations, can also be expressed in terms of the bulk compositions of these sediments. The ratio of feldspar to quartz+quartzite+chert, hereafter termed the $F/Q$ ratio, is a common maturity index for sands (Russell, 1937; Pettijohn, 1949, p. 383). The amount of alkali feldspars (potash feldspars and albite) in percent of total feldspars, hereafter designated as the $(Kf+Ab)/F$ ratio, also measures differences in maturity of the Recent sands along the northern shores of the Gulf of Mexico.

Table 6 lists the average values[2] of various ratios of feldspar and quartz contents. Sands of the eastern Gulf province and of the Sargent and Matagorda beaches contain very small amounts of feldspars; they are orthoquartzitic sands (fig. 6). The sands of the Mississippi River province, on the other hand, contain enough feldspars to form feldspathic sandstones. The sands of the east Texas beaches are transitional in nature. The differences in maturity determined by relative feldspar abundances agree fully with those obtained from heavy-mineral investigations through the use of $S/U$ ratios.

The $(Kf+Ab)/F$ ratio of the sands of the Mississippi River province increases consistently westward, although the amount of alkali feldspars remains almost unchanged. The amount of soda-lime feldspars decreases rapidly from 10.8 percent in Grand Isle sands to about 2 percent in east Texas sands.

## Mineralogy of Recent River Sands

*Heavy-mineral composition of Recent river sands.*—The heavy-mineral suites of the Apalachicola and Mobile river sands are composed almost exclusively of opaque minerals and stable nonopaque heavy minerals such as zircon, tourmaline, staurolite, and kyanite. Hornblende and garnet are present in very small amounts in Apalachicola River sands. The heavy-mineral composition of these river sands is, therefore, very similar to that of the offshore marine sands of the eastern Gulf province.

The Mississippi River sands are characterized by an abundance of unstable heavy minerals such as pyroxenes, amphiboles, and carbonates. Among the stable nonopaque heavy minerals, garnet, epidote, and zircon are common, whereas tourmaline, staurolite, and kyanite are present only in very small amounts (Russell, 1937). The immaturity of the heavy-mineral composition of the Mississippi River sands is similar to that of the offshore marine sands of the Mississippi River province.

[2] Values of individual samples are shown graphically in figure 7.

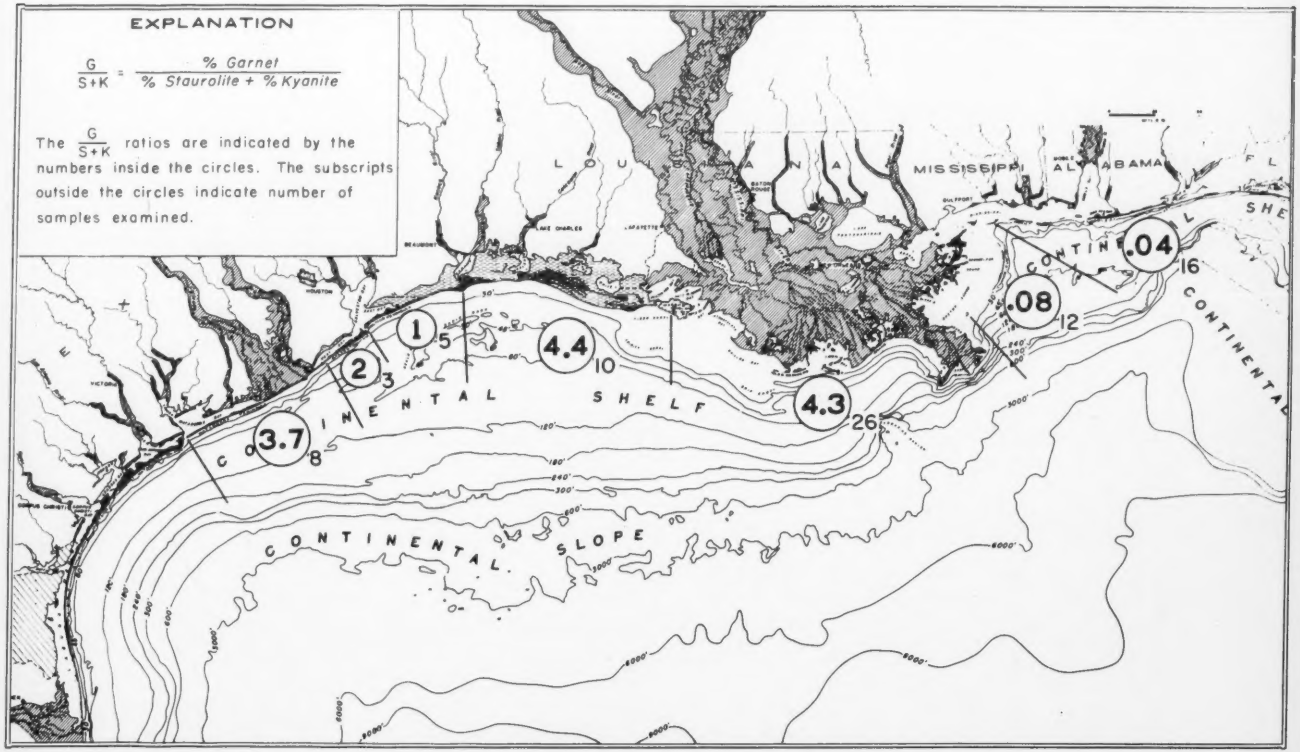


FIG. 5.—Average $G/(S+K)$ ratios in the sands of the Gulf Coast.

TABLE 6.—*Feldspar abundance of Gulf Recent beach sands*

| Locality | No. of Samples | Feldspar Abundance* | | | |
|---|---|---|---|---|---|
| | | $F$ | $F/Q$ | $Kf+Ab$ | $\frac{Kf+Ab}{F}$ |
| (I) *a*) Eastern Gulf | 13 | $\ll 1$ | $\ll 1$ | | |
| (II) Miss. R. Province | | | | | |
| *a*) Delta (Grand Isle) | 8 | 18.7 | 26.2 | 6.8 | 36 |
| *b*) S. La. (Pecan Isle, Cameron Beach) | 4 | 17.0 | 26.3 | 7.7 | 47 |
| *c*) SW La. (Holly Beach) | 3 | 12.1 | 14.2 | 7.6 | 62 |
| (III*a*) East Texas | | | | | |
| *a*) Bolivar | 8 | 7.7 | 8.5 | 5.6 | 73 |
| *b*) Galveston | 4 | 8.6 | 9.7 | 6.4 | 75 |
| *c*) San Luis Beach | 2 | 7.7 | 8.7 | 5.7 | 73 |
| (III*b*) Sargent-Matagorda Beaches | 2 | 4.3 | 4.4 | 2.2 | 50?** |

* $F$ = % Feldspar/total sand ((by volume)
$F/Q$ = number of grains of feldspar per 100 grains of quartz+quartzite+chert
$Kf+Ab$ = % Alkali feldspars/total sand (by volume)
$\frac{Kf+Ab}{F}$ = Alkali feldspar/total feldspars.

** Data are not sufficient to obtain a representative $\frac{Kf+Ab}{F}$ ratio.

The heavy-mineral composition of the sands of the Neches and Trinity Rivers is similar in maturity to that of the river sands of the eastern Gulf Coast states (Bullard, 1942); the nonopaque heavy minerals are predominantly zircon and staurolite. The Brazos and Colorado Rivers, on the other hand, contain abundant hornblende and garnet (Bullard, 1942).

*Feldspar abundance of Recent river sands.*—The sands of the Mississippi, Colorado, and Brazos Rivers differ in maturity from those of the smaller rivers of the Gulf Coastal Plains (fig. 5). The former are arkosic or feldspathic, whereas the latter are orthoquartzitic. The feldspathic sands of the Mississippi River can further be distinguished from the feldspathic sands of the Brazos and Colorado Rivers by the $(Kf+Ab)/F$ ratios; the feldspars are dominantly oligoclase in Mississippi River sands, whereas alkali feldspars predominate in Brazos and Colorado River sands (table 7).

### Mineralogy of Pleistocene Sands

Feldspar abundance of the Pleistocene sands is shown in table 8. The Pleistocene sands of the eastern Gulf are orthoquartzitic, as are the Pleistocene terrace deposits of the Calcasieu and Sabine Rivers of western Louisiana. The Pleistocene deposits of Texas vary somewhat in feldspar abundance. A beach cobble, probably representing reworked Pleistocene Trinity River deposits, contains only 1 percent feldspar. On the other hand, Pleistocene sands from an offshore boring south of Sabine Pass contain 13 percent feldspar. The Pleistocene Mississippi River sands contain as much feldspar as those of today (H. A. Bernard, personal communication). The heavy-mineral composition of the Pleistocene Mississippi River sands is also identical to that of the Mississippi River sands of today (Jones and others, 1954, p. 82–86).

## CHEMICAL COMPOSITION OF RECENT SANDS

### Chemical Composition of Recent Beach Sands

*Variation of the chemical composition of beach sands along the Gulf Coast.*—In the search for methods easier and more rapid than those provided by petrographic tech-

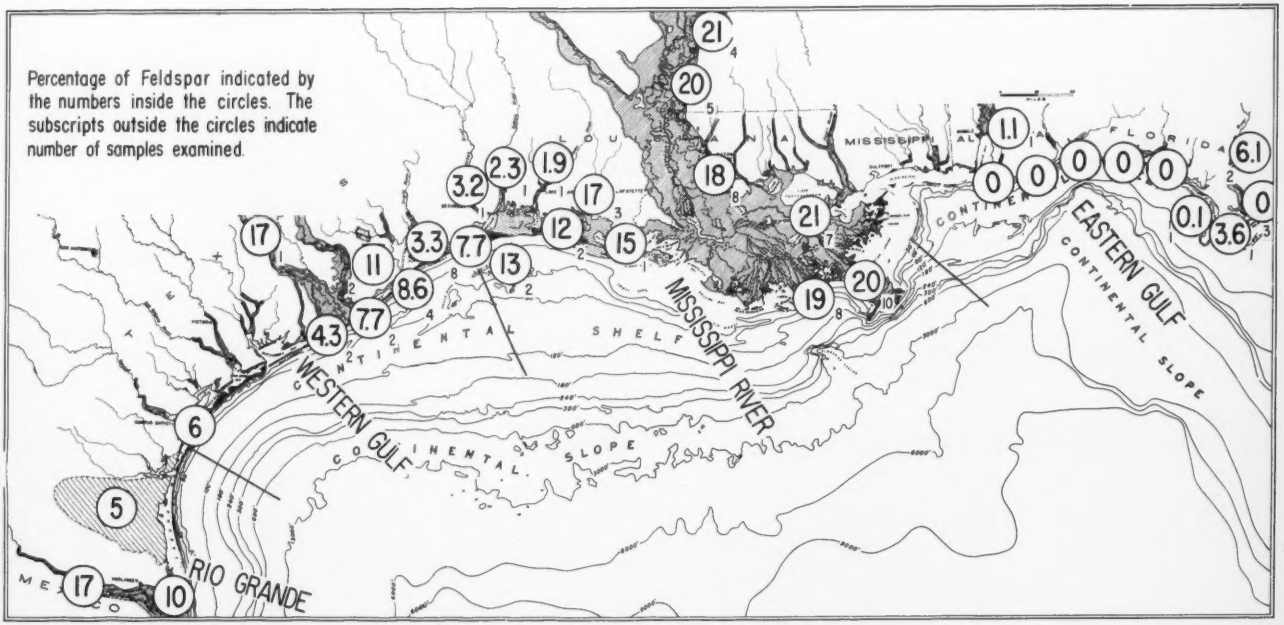


FIG. 6.—Feldspar abundance of Recent sands of the Gulf Coast.

TABLE 7.—*Feldspar abundance of Recent river sands of the Gulf Coast*

| Locality | No. of Samples | Feldspar Abundance* | | | |
|---|---|---|---|---|---|
| | | *F* | *F/Q* | *Kf+Ab* | $\frac{Kf+Ab}{F}$ |
| Eastern Gulf | | | | | |
| Apalachicola River | 2 | 6.1 | 6.8 | 5.5 | 89 |
| Mobile River | 1 | 1.1 | 1.2 | 0.5 | |
| Mississippi River** (mi below Cairo) | | | | | |
| 0– 100 | 4 | 26 | 39 | 7 | 26 |
| 100– 200 | 5 | 25 | 37 | 8 | 31 |
| 200– 300 | 5 | 24 | 37 | 6 | 24 |
| 300– 400 | 5 | 22 | 34 | 8 | 36 |
| 400– 500 | 5 | 22 | 34 | 6 | 26 |
| 500– 600 | 3 | 23 | 34 | 7 | 30 |
| 600– 700 | 5 | 21 | 31 | 7 | 36 |
| 700– 800 | 5 | 20 | 29 | 6 | 30 |
| 800– 900 | 8 | 18 | 25 | 7 | 36 |
| 900–1000 | 7 | 21 | 29 | 8 | 40 |
| 1000–Gulf of Mexico | 10 | 20 | 30 | 7 | 37 |
| Small Rivers of Texas Coastal Plains | 4 | 2.5 | 2.6 | 1.8 | 71 |
| Brazos | 2 | 11.3 | 13.3 | 7.8 | 75 |
| Colorado | 1 | 18.5 | 26.9 | 15.0 | 81 |

* $F$ = % Feldspar/total sand (by volume)
$F/Q$ = number of grains of feldspar per 100 grains of quartz+quartzite+chert
$Kf+Ab$ = % Alkali feldspars/total sand (by volume)
$\frac{Kf+Ab}{F}$ = Alkali feldspar/total feldspars.
** Averages, computed from the data published by Russell (1937).

niques for determining the mineral composition of sands, spectrochemical techniques were tried. The degree of maturity of Gulf Recent sands, as revealed by mineralogical studies, is manifested by the differences in chemical composition. The high $K_2O$ and $Na_2O$ contents of the Recent sands of the Mississippi River province indicate abundant feldspar, and the high CaO, MgO, and $Fe_2O_3$ contents suggest a concentration of heavy minerals. If we assume that all $K_2O$, $Na_2O$, and CaO in the samples are combined

TABLE 8.—*Feldspar abundance of Pleistocene sands of the Gulf Coast*

| Locality | No. of Samples | Feldspar Abundance* | | | |
|---|---|---|---|---|---|
| | | *F* | *F/Q* | *Kf+Ab* | $\frac{Kf+Ab}{F}$ |
| Eastern Gulf | 2 | 0 | 0 | 0 | |
| Small river terrace deposits | 5 | Rare | — | Rare | |
| Marine sands off Sabine Pass from boring | 2 | 12.9 | 16.1 | 10.1 | 80 |
| Barrier beach sands at Smith Point | 1 | 3.3 | 3.5 | 2.9 | 88 |
| Beach pebble from Bolivar Beach | 1 | 1.3 | 1.3 | 0.4 | — |
| Beach cobble from Sargent Beach | 1 | 12.0 | 13.7 | 9.3 | 78 |

* $F$ = % Feldspar/total sand (by volume)
$F/Q$ = number of grains of feldspar per 100 grains of quartz+quartzite+chert
$Kf+Ab$ = % Alkali feldspars/total sand (by volume)
$\frac{Kf+Ab}{F}$ = Alkali feldspar/total feldspars.

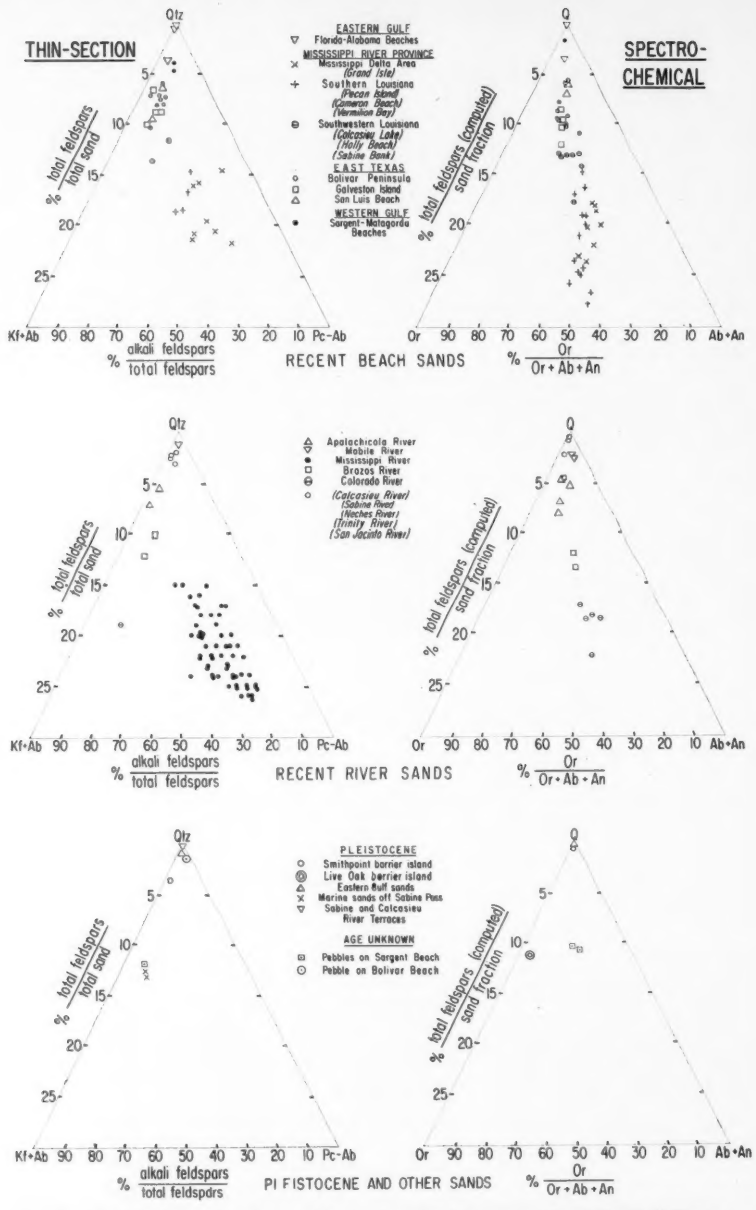


FIG. 7.—Feldspar abundance of Recent and Pleistocene sands of the Gulf Coast.

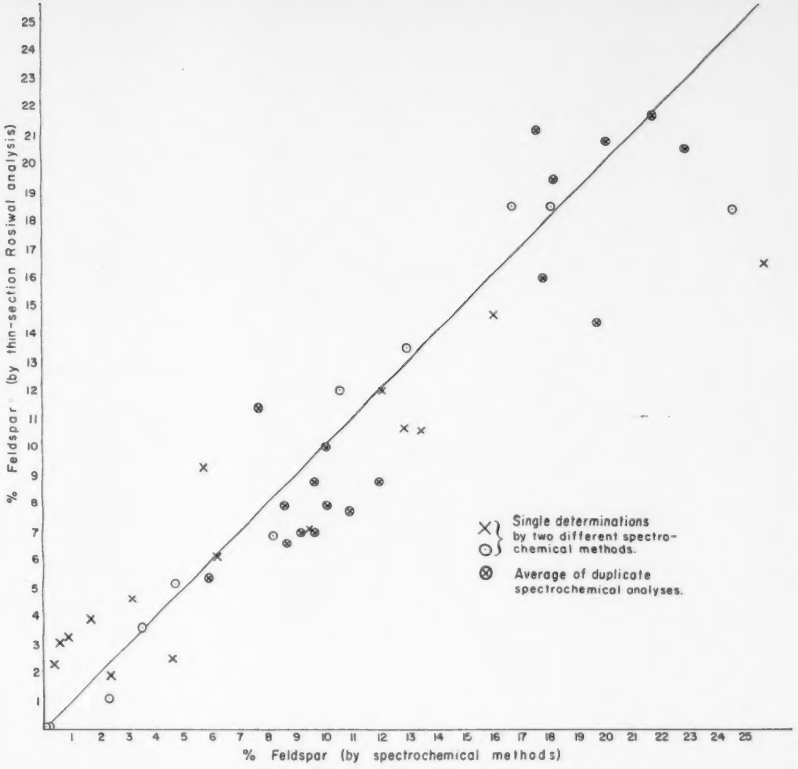
% Feldspar (by thin-section Rosiwal analysis)
% Feldspar (by spectrochemical methods)
Single determinations by two different spectrochemical methods.
Average of duplicate spectrochemical analyses.

FIG. 8.—Correlation of feldspar abundance computed from spectrochemical analysis and feldspar abundance determined by thin-section analysis.

with $Al_2O_3$ and $SiO_2$ to form feldspars, the spectrochemical results can be calibrated to the petrographic results. The computed values of feldspar abundance (fig. 7) are, on the average, slightly higher than those determined petrographically; the correlation of spectrochemical and petrographic results is shown in figure 8.

The $Or/F$ (orthoclase/total feldspar) ratio, though not identical to the $(Kf+Ab)/F$ ratio, serves as a measure of the relative abundance of potassium and plagioclase feldspars. A change in $Or/F$ ratios of the Gulf Recent sands from the Mississippi River Delta westward is noted, but it is far less significant than the corresponding change in $(Kf+Ab)/F$ ratios (fig. 7).

The differences in MgO content of the Gulf Recent sands are significant. Amphiboles, pyroxenes, and garnets are the most common magnesium-bearing, acid-insoluble heavy minerals in these sands. The decrease in MgO from the Mississippi River Delta area westward (table 9) indicates a decreasing abundance of these minerals. This decrease is not necessarily a decrease in relative abundance within the heavy mineral suite. In fact, the relative amount of amphibole, which is the most important magnesium-bearing heavy mineral, in percent of total heavy minerals, remains practically unchanged from the Mississippi Delta area to east Texas (Goldstein, 1942). Apparently, there is a decrease of all heavy minerals west

TABLE 9.—*Average chemical compositions of Recent beach sands of the Gulf Coast (in percent)*

| Locality | No. of Samples | $K_2O$ | $Na_2O$ | CaO | MgO | $Fe_2O_3$ | $Al_2O_3$ | $SiO_2$ |
|---|---|---|---|---|---|---|---|---|
| Eastern Gulf Beaches | | | | | | | | |
| 1) Interdeltaic | 22 | 0.007 | 0.088 | 0.06 | 0.02 | 0.08 | 0.027 | 99.72 |
| 2) Apalachicola Delta | 1 | 0.32 | 0.25 | 0.10 | 0.02 | 0.07 | 0.80 | 98.44 |
| Mississippi River Province | | | | | | | | |
| 1) Delta Area | 8 | 1.3 | 1.2 | 1.3 | 0.54 | 1.2 | 6.6 | 87.86 |
| 2) S. Louisiana | 16 | 1.7 | 1.3 | 0.46 | 0.34 | 1.2 | 5.5 | 89.50 |
| 3) SW Louisiana | 7 | 1.1 | 0.71 | 0.23 | 0.12 | 0.36 | 3.6 | 93.88 |
| East Texas Beaches | | | | | | | | |
| 1) Bolivar Beach | 10 | 0.79 | 0.39 | 0.16 | 0.05 | 0.28 | 1.9 | 96.43 |
| 2) Galveston Island | 4 | 0.92 | 0.44 | 0.20 | 0.12 | 0.40 | 2.5 | 95.42 |
| 3) San Luis Beach | 2 | 0.15 | 0.30 | 0.07 | 0.05 | 0.18 | 1.4 | 97.49 |
| Western Gulf Beaches | | | | | | | | |
| Sargent-Matagorda | 2 | 0.23 | 0.11 | 0.05 | 0.04 | 0.12 | 0.42 | 99.03 |

of the Mississippi River Delta. This decrease is also indicated by the differences in $Fe_2O_3$ content of the Gulf Recent sands of the Mississippi River Delta. The sands of the eastern Gulf province contain very little garnet, amphibole, and pyroxene, and their MgO content is very low.

The $Al_2O_3$ and $SiO_2$ contents of sands can also be used as maturity indices, as indicated by the differences in Gulf Recent sands of different degrees of maturity (table 9).

*Chemical composition of the various size fractions of beach sands.*—The chemical composition of the various size fractions of the Bolivar (Texas) beach sands was determined spectrochemically. Two groups of samples were analyzed. Group I comprises five samples from localities 1610 and 1611 on the eastern half of the Bolivar beach, near the mouth of the Sabine River. Group II comprises four samples from localities 1605 and 1606 on the western half of the Bolivar beach. Table 10 shows the average chemical composition of the seven sieve fractions of these two groups of samples. Particularly remarkable is the systematic decrease of the $K_2O$ and $Na_2O$ contents with grain size in the sands from group I. Group II sands also show a decrease of $K_2O$ with grain size, but the variation of $Na_2O$ is insignificant. Variations in $K_2O$ and $Na_2O$ contents are believed to reflect changes in feldspar abundance. On the assumption that all $K_2O$, $Na_2O$, and CaO are combined with $Al_2O_3$ and $SiO_2$ to form

TABLE 10.—*Average chemical composition of the various size fractions of the Bolivar Beach sands*

| | | No. of Samples | $K_2O$ | $Na_2O$ | CaO | MgO | $Fe_2O_3$ | $Al_2O_3$ | $SiO_2$ |
|---|---|---|---|---|---|---|---|---|---|
| Group I | 0.250 mm fraction | 5 | 0.89 | 0.73 | 0.17 | 0.13 | 16.0 | 2.8 | 79.5 |
| | 0.210 mm fraction | 5 | 1.2 | 0.79 | 0.13 | 0.03 | 0.49 | 1.8 | 95.6 |
| | 0.177 mm fraction | 5 | 1.1 | 0.66 | 0.13 | 0.03 | 0.31 | 2.0 | 95.8 |
| | 0.149 mm fraction | 5 | 0.87 | 0.54 | 0.13 | 0.02 | 0.22 | 1.5 | 96.7 |
| | 0.125 mm fraction | 5 | 0.65 | 0.45 | 0.13 | 0.03 | 0.24 | 1.8 | 96.7 |
| | 0.088 mm fraction | 5 | 0.45 | 0.37 | 0.13 | 0.03 | 0.26 | 1.4 | 97.3 |
| | 0.062 mm fraction | 5 | 0.60 | 0.60 | 0.18 | 0.22 | 3.0 | 2.0 | 93.4 |
| Group II | 0.250 mm fraction | 4 | 1.1 | 0.61 | 0.22 | 0.08 | 3.1 | 3.2 | 91.6 |
| | 0.210 mm fraction | 4 | 1.1 | 0.66 | 0.22 | 0.06 | 0.49 | 3.2 | 94.3 |
| | 0.177 mm fraction | 4 | 1.0 | 0.62 | 0.22 | 0.06 | 0.39 | 2.7 | 95.0 |
| | 0.149 mm fraction | 4 | 0.94 | 0.60 | 0.21 | 0.05 | 0.31 | 2.4 | 95.5 |
| | 0.125 mm fraction | 4 | 0.82 | 0.66 | 0.18 | 0.05 | 0.27 | 2.6 | 95.5 |
| | 0.088 mm fraction | 4 | 0.73 | 0.63 | 0.17 | 0.07 | 0.39 | 2.4 | 95.6 |
| | 0.062 mm fraction | 4 | 0.76 | 0.81 | 0.38 | 0.27 | 2.5 | 3.4 | 91.9 |

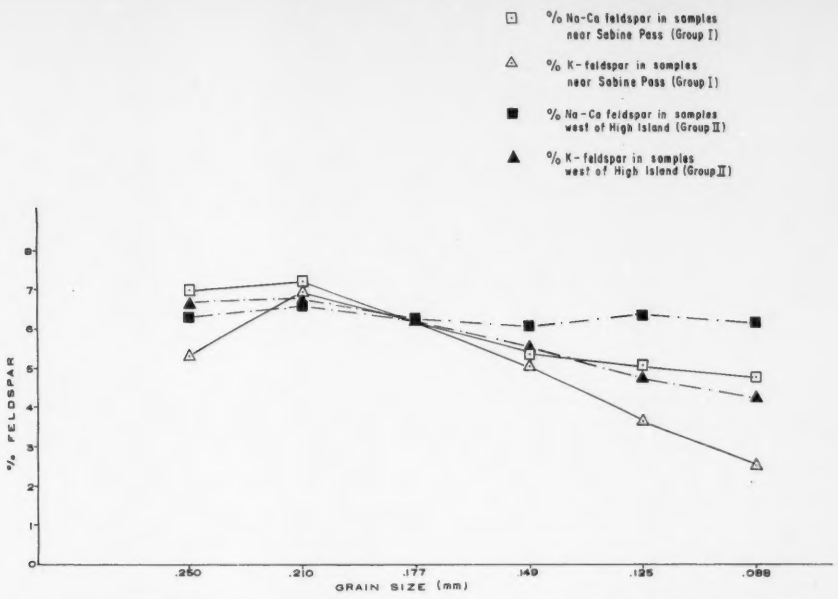


FIG. 9.—Computed feldspar abundance of the various size fractions of the Bolivar Beach sands.

feldspars, the feldspar abundance of the various size fractions of the Bolivar sands has been computed from spectrochemical results; the variation of average feldspar abundance with grain size for the sands from groups I and II is shown in figure 9. The feldspar abundance of the size fraction between 0.088 and 0.062 mm has not been computed, because binocular examinations revealed an abundance of micas in this finest fraction; the micas should account for a considerable part of the alkali content in this fraction.

The $Fe_2O_3$ content of the coarsest fraction of the Bolivar beach sands is exceptionally high. Binocular examinations revealed the presence of numerous limonite particles, which presumably represent a part of the cementing materials for the offshore sediments.

### Chemical Composition of Recent River Sands and Pleistocene Sands

Sand samples from the Mobile, Calcasieu, Sabine, Neches, Trinity, and San Jacinto Rivers all contain very little $K_2O$ and $Na_2O$, as is to be expected from their mineralogical compositions (table 11). Forty-one samples of Recent and Pleistocene sands, collected from the Pearl, Tangipahoe, Amite, Calcasieu, Sabine, and Red Rivers and from small tributaries to the Mississippi River, were chemically analyzed by the Louisiana Geological Survey (Woodward and others, 1941); these sands are also very low in alkali content.

The Apalachicola, Brazos, and Colorado River sands contain more $K_2O$ and $Na_2O$ than the sands from the smaller rivers. No chemical analyses of the Mississippi River sands are available.

The feldspar abundance of the Recent river sands of the Gulf Coast, as computed from the spectrochemical results, is shown in figure 7.

## GENESIS OF THE RECENT SANDS

### River Sands

The composition of the river sands depends primarily upon the source materials. Rivers that have their headwaters in igneous

TABLE 11.—*Average chemical composition of Recent River sands and Pleistocene sands of the Gulf Coast (in percent)*

| Locality | No. of Samples | $K_2O$ | $Na_2O$ | CaO | MgO | $Fe_2O_3$ | $Al_2O_3$ |
|---|---|---|---|---|---|---|---|
| Apalachicola River | 4 | 0.68 | 0.26 | 0.19 | 0.06 | 0.26 | 1.8 |
| Mobile River | 2 | 0.18 | 0.18 | 0.08 | 0.02 | 0.14 | 1.0 |
| San Jacinto, Trinity, Neches, Sabine, Calcasieu Rivers | 4 | 0.19 | 0.046 | 0.025 | 0.014 | 0.16 | 0.36 |
| Brazos River | 2 | 1.1 | 0.73 | 0.10 | 0.10 | 0.37 | 2.4 |
| Colorado River | 5 | 1.6 | 0.95 | 0.25 | 0.20 | 0.48 | 3.9 |
| Pleistocene, barrier islands | | | | | | | |
| Smith Point Bar | 1 | 0.10 | 0.029 | 0.013 | 0.01 | 0.04 | 0.22 |
| Live Oak Bar | 1 | 1.6 | 0.18 | 0.24 | 0.03 | 0.15 | 1.6 |
| Pleistocene, Eastern Gulf | 2 | 0.009 | 0.059 | 0.07 | 0.02 | 0.10 | 0.027 |
| Beach cobble on Sargent Beach | 2 | 0.88 | 0.50 | 0.20 | 0.12 | 0.28 | 3.4 |

and metamorphic rocks, such as the Mississippi, Apalachicola, Colorado, and Rio Grande, carry significant amounts of feldspar and unstable heavy minerals. Smaller streams on the coastal plains and tributaries to the lower Mississippi River carry only second-cycle, or even polycycle, sediments; accordingly, these river sands contain very little feldspar and relatively stable heavy minerals.

### Beach Sands East of the Mississippi River Delta

The Apalachicola River is the largest river on the eastern Gulf Coastal Plains. The beach sands of the eastern Gulf, however, are coarser grained and contain much less feldspar than the sands of the Apalachicola River. Only the sand immediately west of the mouth of the Apalachicola on Indian Lagoon beach is similar in mineralogical composition to the Apalachicola River sand. Less than 5 miles west on Cape San Blas beach, the sand again contains practically no feldspar and, in this respect, is similar to other beach sands in the interdeltaic areas of the eastern Gulf. Most of the Recent beach sands are similar to the Pleistocene sands in grain size and mineralogy. Thus, it is probable that the beach sands of northwest Florida were derived by the reworking of older (mainly Pleistocene) sediments and that the modern Apalachicola River has not contributed much material to the beaches. The Pleistocene sands of the eastern Gulf were probably derived from the early Tertiary formations, with minor contributions from rocks of the southern Appalachians. As the early Tertiary rocks of the southeastern states are poor in feldspar and unstable heavy minerals (Grim, 1936), it is not surprising that the eastern Gulf Recent sands contain practically no feldspars and are characterized by a very mature heavy-mineral suite.

### Beach Sands in the Mississippi River Delta Area

The beach sand on Grand Isle in the Mississippi River Delta area is similar to Mississippi River sand in almost every aspect—grain size, sorting, bulk composition, feldspar abundance [including both $F/Q$ and $(Kf+Ab)/F$ ratios], and heavy-mineral composition. The source material for the beach sands in the Mississippi River Delta area is furnished essentially by the Mississippi River.

### Beach Sands West of the Mississippi River Delta

The coastal plains of southwestern Louisiana and southeastern Texas are drained by small consequent streams, including the Calcasieu, Sabine, Neches, Trinity, and San Jacinto Rivers. The river sands contain only about 2–3 percent feldspars. The sands on the beaches of the region, on the other hand, contain three or four times as much feldspar. Thus, the source material for these beaches must have been at least partly furnished by the Mississippi River, which carries a feldspathic sand.

Differences in mineralogical and chemical

compositions between the Mississippi River sands and the east Texas beach sands are shown by table 6 and by figure 10. The Mississippi River sand is feldspathic; it contains about 20 percent feldspars, dominantly soda-lime feldspars. The east Texas beach sands are mostly orthoquartzitic. Such differences in composition may be explained by one or a combination of the following hypotheses:

1.—The beach sands of the Mississippi Delta area and of Texas were both derived from the Mississippi River. The difference in feldspar abundance between the two resulted from a preferential elimination of the feldspars during the westward transport of the Mississippi River sands to Texas beaches.

2.—The beach sands of the Mississippi Delta area and of Texas were both derived from the Mississippi River. The difference in feldspar abundance between the two resulted from a preferential elimination of the feldspars during the deposition of the Texas beach sands, which were less quickly buried and more thoroughly winnowed than the beach sands of the Mississippi Delta area.

3.—The beach sands of the Mississippi Delta area were derived from the Mississippi River. The Texas beach sands, however, were derived partly from the Mississippi River, partly from the coastal rivers, and partly from the offshore erosion of Pleistocene and earlier Recent deposits. The difference in feldspar abundance between the beach sands in the Mississippi Delta area and those in Texas resulted from a mixing of the Mississippi River sands with feldspar-poor sands from other sources.

*Evaluation of the possible effect of abrasion and solution.*—As feldspars are less resistant to abrasion than quartz, they may have beeen differentially eliminated during the westward transportation of the Mississippi River sands by longshore currents.

Fractional loss of a feldspar by abrasion is essentially a process of wearing feldspar grains down to clay-size particles; it is therefore a direct function of the distance traveled but is independent of the amount of the feldspar present. If (1) there is no loss of feldspar during transport, and (2) there is no contamination by other sources, and (3) if losses of feldspar are due entirely to abrasion and solution during transport by longshore currents, the amount of a feldspar at any given point, $f$, would be related to that at Grand Isle, $f_0$ by a logarithmic function (compare Plumley, 1948):

$$f=f_0e^{-kd}, \tag{1}$$

where $d$ is the distance from Grand Isle, and $k$ is a constant related to $d_{1/2}$ (distance required to reduce the feldspar by half of the amount present) by the relation

$$d_{1/2}=\frac{\log_e 2}{k}. \tag{2}$$

If the east Texas sands were derived entirely from the Grand Isle sands, such a relation should not exist. In fact, such a relation does not exist. The following observations are particularly significant:

1.—The Grand Isle sand contains 18.7 percent total feldspars and 11.9 percent soda-lime feldspars; the Cameron beach sand contains 17.8 percent total feldspars and 9.3 percent soda-lime feldspars. After 180 miles of travel, the decrease in feldspar abundance is almost negligible.

2.—Holly beach is only about 10 miles west of the Cameron beach, but the Holly

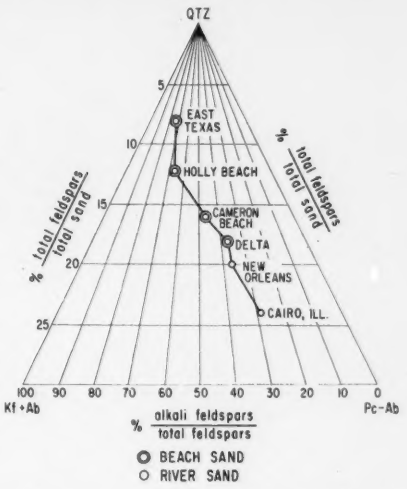


FIG. 10.—Variation of feldspar abundance of Recent sands from Cairo, Illinois to East Texas beaches.

beach sand contains only one-half as much soda-lime feldspars and about two-thirds as much total feldspars as the Cameron beach sand.

3.—The Bolivar and Galveston beaches extend continuously for almost 100 miles from near Sabine Pass to San Luis Pass. Thin-section and chemical analyses of closely spaced samples show no systematic westward change in sand composition from Sabine Pass to San Luis Pass. On the contrary, the Galveston samples commonly contain more (approximately 1 percent) feldspars per 100 quartz minerals than the Bolivar samples.

Thus, it seems that the loss of feldspars by abrasion during transport by longshore currents has not been significant enough to outweigh all other factors and produce a systematic westward change in sand composition.

*Evaluation of the possible combined effect of abrasion and size sorting.*—The total feldspar content of the Mississippi River sand is greater in finer size fractions (Russell, 1936); a breakdown of coarser feldspar particles due to abrasion might have resulted in a concentration of feldspars in the finer grades. As beach sediments of finer grades can more easily be carried offshore by waves and undertows than those of coarser grades, a preferential removal of the finer-grade particles would result in a decrease of the feldspar abundance of the sand as a whole. Thus, it is possible that abrasion and solution combined with size sorting could have produced the decrease in feldspar abundance of the beach sands west of the Mississippi River Delta.

If this process had indeed taken place, the finer fractions of beach sands should be richer in feldspars than the coarser fractions. Spectrochemical analysis of the feldspar content of the various size fractions of the Bolivar beach sands shows that the reverse is true (fig. 9). Thus, the westward decrease of feldspars on the Gulf Coast beaches could not have resulted simply from a combined effect of abrasion and size sorting.

*Evaluation of the possible effect of dilution on the basis of feldspar abundance.*—If we assume that loss of feldspars during transportation and after deposition is minor, the relative importance of different sources for the beach sands of southwestern Louisiana and east Texas can be estimated as follows.

TABLE 12.—*Petrographically determined feldspar abundance of Holly Beach sand (in percent)*

| | Alkali Feldspars | Soda-Lime Feldspars | Quartz, Etc. |
|---|---|---|---|
| Sand I | 8 | 9 | 85 |
| Sand II | 10 | 3 | 88 |
| Sand III | 2 | 0.5 | 97 |

## Holly Beach, Louisiana

The possible sources for the Holly beach sands can be grouped into three categories:

Sand I—Sands swept westward by longshore currents, consisting chiefly of (a) sands carried down by the present Mississippi River, and (b) sands carried down by the early Recent Mississippi River, laid down as deltaic or offshore marine deposits, and now being reworked.

Sand II—Sands carried onshore by waves, consisting chiefly of (a) sands carried down by the Pleistocene Mississippi River, laid down as offshore marine sands, and now being reworked, and (b) sands of unknown source on Sabine Bank.

Sand III—Sands carried down by small, coastal-plain rivers of Recent and/or Pleistocene age.

With the average composition of the beach sands of southern Louisiana being taken as the composition of sand I, the average composition of the Pleistocene offshore marine sands as that of sand II, and the average composition of the sands of the small, coastal-plain rivers as that of sand III (table 12), the relative abundance of sands of these three sources on Holly beach can be computed as follows:

$$8\%a+10\%b+2\%c=7.6\%$$
$$9\%a+3\%b+0.5\%c=4.5\%$$
$$a+b+c=100\%.$$

Solution of these equations gives

$$a=\%\text{ sand I on Holly beach} \cong 35\%$$
$$b=\%\text{ sand II on Holly beach} \cong 45\%$$
$$c=\%\text{ sand III on Holly beach} \cong 20\%.$$

The conclusion reached from the analysis of the feldspar-composition data of the sands

TABLE 13.—*Computed and spectrochemically determined chemical composition in Holly Beach sand (in percent)*

| | $K_2O$ | $Na_2O$ | CaO | MgO | $Fe_2O_3$ | $Al_2O_3$ | $SiO_2$ | Fraction of Sand |
|---|---|---|---|---|---|---|---|---|
| Sand I | 1.7 | 1.3 | 0.46 | 0.34 | 1.2 | 5.5 | | 35 |
| Sand II | 1.2 | 0.54 | 0.26 | 0.09 | 0.24 | 4.2 | | 45 |
| Sand III | 0.2 | 0.05 | 0.025 | 0.014 | 0.16 | 0.36 | | 20 |
| Holly Beach Comp'd | 1.2 | 0.71 | 0.28 | 0.16 | 0.57 | 3.9 | 93.2 | |
| Holly Beach Det'd | 1.1 | 0.78 | 0.22 | 0.13 | 0.46 | 3.4 | 93.9 | |

can be checked by an analysis of the chemical-composition data. Table 13 lists the analytically determined average chemical composition as compared with the theoretically computed chemical composition of the Holly beach sand. The agreement is so close that the general validity of the estimate is probable.

Thus, if losses due to abrasion are considered negligible, the following conclusions can be drawn:

1.—About 35 percent of the sand on Holly beach is derived from the present and the ancestral Mississippi River, and the small, coastal-plain rivers supply only a small portion (perhaps only 20 percent) of source material to Holly beach.

2.—Although longshore currents play an important role in transporting the Mississippi River sands westward, offshore marine erosion has also supplied much material (45 percent) to Holly beach.

### Bolivar and Galveston Beaches, Texas

The possible sources of the sands on the east Texas beaches can be grouped into three categories:

Sand A—Sands swept westward by longshore currents, consisting chiefly of sands carried down by the Mississippi River and diluted by sands from other sources during transportation.

Sand B—Sands swept westward by longshore currents, consisting chiefly of sands carried down by the Pleistocene Mississippi River and now being reworked.

Sand C—Sands carried down by small, coastal-plain rivers of Recent and/or Pleistocene age.

If the average composition of the Holly beach sands is taken as the composition of sand A, the average composition of the Pleistocene offshore marine sands as that of sand B, and the average composition of the sands of the small, coastal-plain rivers as that of sand C (table 14), and if the analytically determined composition of the beach sands on the east Texas coast (5.8 percent alkali feldspars and 2.2 percent soda-lime feldspars) is used, the relative abundance of sands of these three sources on east Texas beaches can be computed as follows:

$$7.6\%a+10\%b+2\%c=5.8\%$$
$$4.7\%a+3\%b+0.5\%c=2.2\%$$
$$a+b+c=100\%.$$

Solution of these equations gives

$$a=\%\text{ sand } A \text{ on east Texas beaches} \cong 20\%$$
$$b=\%\text{ sand } B \text{ on east Texas beaches} \cong 35\%$$
$$c=\%\text{ sand } C \text{ on east Texas beaches} \cong 45\%.$$

The conclusion reached from the analysis of the feldspar-composition data of the sands can be checked by an analysis of the chemical-composition data. Table 15 lists the experimentally determined average chemical composition of the east Texas beach sands as compared with the theoretically computed chemical composition. The agreement is so close that the general validity of the estimate is probable.

Thus, if losses due to abrasion and sorting

TABLE 14.—*Petrographically determined feldspar abundance in East Texas Beach sands (in percent)*

| | Alkali Feldspars | Soda-Lime Feldspars | Quartz, Etc. |
|---|---|---|---|
| Sand A | 7.6 | 4.7 | 88 |
| Sand B | 10 | 3 | 87 |
| Sand C | 2 | 0.5 | 97 |

TABLE 15.—*Computed and spectrochemically determined chemical composition of East Texas beach sands** (*in percent*)

| | $K_2O$ | $Na_2O$ | CaO | MgO | $Fe_2O_3$ | $Al_2O_3$ | $SiO_2$ | Fraction of Sand |
|---|---|---|---|---|---|---|---|---|
| Sand A | 1.1 | 0.78 | 0.22 | 0.13 | 0.48 | 3.4 | | 20 |
| Sand B | 1.2 | 0.54 | 0.26 | 0.09 | 0.24 | 4.2 | | 35 |
| Sand C | 0.2 | 0.05 | 0.025 | 0.014 | 0.16 | 0.36 | | 45 |
| East Texas Beach Comp'd | 0.73 | 0.37 | 0.16 | 0.04 | 0.25 | 2.3 | 96.1 | |
| East Texas Beach Det'd | 0.85 | 0.42 | 0.18 | 0.09 | 0.34 | 2.2 | 95.9 | |

* Average composition of Galveston and Bolivar beach sands.

are negligible, the following conclusions can be drawn:

1.—Approximately 45 percent of the source material for east Texas beaches was supplied by small, coastal-plain rivers, probably through reworking of Pleistocene deposits.

2.—Although longshore currents play an important part in transporting sediments westward, offshore marine erosion has supplied considerable material to the east Texas beaches.

San Luis, Sargent, and Matagorda Beaches

Data on the San Luis, Sargent, and Matagorda beaches are too meager to permit an estimate of the relative abundance of sands from different sources. The fact that the Colorado and Brazos River sands are feldspathic whereas the beach sands are orthoquartzic suggests that offshore marine erosion of orthoquartzitic sediments supplies the bulk of material to these beaches. The Colorado River has only recently been diverted to its present position; perhaps the bulk of the feldspathic sands carried down by the river during the Recent has been swept westward by longshore currents and has been deposited on the barrier beaches west of the present Colorado River Delta on Matagorda Island (Bullard, 1942).

*Evaluation of the possible effect of dilution on the basis of the size distribution of feldspars.*—The fact that the sand on the Texas beaches is partly derived from coastal-plain rivers is suggested by the size variations of the beach sand. Sands carried to the coast by small, coastal-plain rivers are, on the average, finer grained than the sand transported along the Texas coast by longshore currents; sands from the San Bernard River Delta, for example, have an average median diameter of 0.090 mm, which is almost 0.050 mm finer than the average beach sand. Mixing of a river sand and that transported by longshore currents resulted in the finer grain size of the beach sand near the mouth of a river (fig. 3).

The sands carried to the coast by the small, coastal-plain rivers, such as the Sabine and Trinity Rivers, are not only finer but are also poor in feldspars. Beach sands resulting from a mixing of the very fine sand derived from those sources with the coarser sand supplied by the Mississippi River and carried westward by longshore currents should be poorer in feldspars in the fine fractions and richer in feldspars in the coarser fractions. The sands on the eastern half of the Bolivar beach are indeed poor in feldspars in the finer fractions (fig. 9), and the variation of feldspar abundance with grain size can be explained if such a mixing has taken place. The amount of mixed river materials is less in the sands on the western half of the Bolivar beach, where the variation of feldspar content with grain size is less significant and the total feldspar content is higher.

*Evaluation of the possible effect of dilution on the basis of heavy minerals.*—Garnet, staurolite, and kyanite have similar specific weights; thus, the $G/(S+K)$ ratio should not be greatly affected by size sorting. Staurolite, kyanite, and garnet also have about the same degree of hardness. Stauro-

lite and kyanite, particularly the latter, possess good cleavage. They are probably less resistant to abrasion than garnets, which have no cleavage. The relative resistance of these minerals to solution during transport is not certain, although they are commonly grouped fairly close together in empirical lists of the order of stability of heavy minerals (Pettijohn, 1949, p. 489). Therefore, it is reasonable to conclude that the combined effect of abrasion and solution during transports should either keep the $G/(S+K)$ ratio approximately unchanged or, more probably, increase the $G/(S+K)$ ratio because of the lesser resistance of staurolite and kyanite to abrasion.

Many outcropping Cretaceous and Tertiary formations of Texas contain a heavy-mineral suite rich in staurolite and kyanite but poor in garnet (Bornhauser, 1940; Cogen 1940). Consequently, second-cycle river sands on coastal plains have $G/(S+K)$ ratios less than 1 (Bullard, 1942). Thus, if the Mississippi River sands transported to the east Texas coast have been diluted by a significant amount of sands carried by small, coastal-plain rivers, the resulting beach sands should have lower $G/(S+K)$ ratios than the sands of the Mississippi River province.

Marine sands of the Mississippi Delta area and offshore from the Louisiana coast have an average $G/(S+K)$ ratio of about 4.5, whereas the average east Texas beach sand has a $G/(S+K)$ ratio of about 1.5 (fig. 5). This difference suggests that dilution far outweighs abrasion and solution in changing the composition of the Gulf Recent beach sands.

*Evaluation of the effect of dilution on the basis of physiographic studies.*—The fact that offshore marine erosion must have contributed to the source materials of the Texas beach sand is suggested by (1) the presence of abundant beach pebbles and cobbles, which represent offshore deposits eroded and carried onshore by waves, and (2) the physiographic history of the beaches. The Bolivar beach, for example, has been considered a transgressive beach resulting from a marine transgression over the Pleistocene Trinity River Delta (H. A. Bernard, personal communication). Marine erosion of this delta must have contributed materials to the

TABLE 16.—*Depositional environments of sedimentary basins of the Gulf Coast*

| | Eastern Gulf Platform | Mississippi River Delta | Western Gulf Platform | | | Rio Grande Delta |
|---|---|---|---|---|---|---|
| | | | East Texas | Western Gulf | South Texas | |
| Source material for Recent River Sands | Mesozoic and Cenozoic sediments of the Gulf Coastal Plain, with minor contribution from southern Appalachian igneous rocks | Igneous metamorphic and sedimentary rocks of the Mississippi River drainage basin | Mesozoic and Cenozoic sediments of the Gulf Coastal Plain | A. Mesozoic and Cenozoic sediments of the Gulf Coast Plain<br>B. Igneous and metamorphic rocks of the Llano region | Mesozoic and Cenozoic sediments of the Gulf Coast Plain | Igneous, metamorphic, and sedimentary rocks of the Rio Grande drainage basin |
| Relief of Source Areas | Low, in general | Considerable in some parts (for example, Rocky Mountains) | Low | Low | Low | Considerable in some parts (for example, southrn Rocky Mountains) |
| Climate of Source Areas | Warm and humid in general | Varied, cold and dry in places | Warm and humid | Warm and dry inland and humid near coast | Warm and dry | Varied, warm and dry in places |
| Environment of Basin of Deposition | Slow deposition of second-cycle sediments | Rapid deposition of sediments carried down by the Mississippi | Slow deposition of sediments carried from delta areas by longshore currents, diluted by reworked, second-cycle sediments | | | Rapid deposition of sediments carried down by the Rio Grande |

beach. As the eroded materials were poor in feldspar, a mixing of these materials with the sand transported westward from the Mississippi River Delta by longshore currents should have resulted in a sand which is poorer in feldspar than the sand of the Mississippi River Delta.

## SUMMARY

The maturity of a sand may depend on (1) the composition of the source materials, (2) abrasion and solution (combined with size sorting) during transport, and (3) abrasion and solution (combined with size sorting) during deposition. The preceding evaluation of the relative importance of the various factors suggests that the composition of the Gulf Recent sands depends primarily on the source materials. Decreases of feldspar abundance of the beach sands west of the Mississippi River Delta can be accounted for by dilution from other sources. Losses of feldspar by abrasion and solution during transport and deposition cannot be evaluated quantitatively, but they are probably minor.

The source materials for the beach sands west of the Mississippi River Delta are partly carried down by the Mississippi River and transported westward by longshore currents, partly carried down by coastal-plain rivers, and partly supplied by offshore erosion of previously deposited coastal-plain or deltaic deposits.

The composition of source materials reflects the relief, climate, and composition of terranes of the source areas. As indicated by table 16, all factors tend to sharpen the difference in composition between the eastern Gulf sand and the beach sand of the Mississippi River Delta area. However, interpretation of the maturity of sediments is complicated by the fact that longshore currents are effective agents in transporting sediments from geosynclinal to stable areas. Although the tectonic framework of the stable western Gulf area is not greatly different from that of the eastern Gulf platform, the beach sands on the Texas coast (despite considerable dilution by second-cycle sediments) are far less mature than the Florida sands because of the prevailing westward direction of the longshore current.

## ACKNOWLEDGMENT

The writer is particularly indebted to R. H. Nanz for his continued assistance in the formulation and execution of the investigation. The advice and guidance given by H. A. Bernard and R. J. LeBlanc are also greatly appreciated. Grateful acknowledgment is due W. A. Begin, J. Laverde, and R. J. Grabowski for their aid in the analysis of the sediments and presentation of the results. The original manuscript has been improved by the valuable constructive criticism of R. H. Nanz, H. A. Bernard, B. W. Wilson, Mrs. J. Breeding, and Miss B. Shearer.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 404–413
FIGS. 1–3, SEPTEMBER, 1960

# MINERALOGY AS AN INDICATOR OF BEACH SAND MOVEMENT ALONG THE RHODE ISLAND SHORE[1]

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## ABSTRACT

Southwestern Rhode Island beach sands are composed of a great variety of minerals, the most common of which are amphiboles, chlorite, garnet, staurolite, and black opaques in the heavy fraction and feldspars and quartz in the light fraction. Counts of these fractions from samples collected at one-mile intervals were used as basic data for multivariate and trend analyses. Character of source materials and pattern of beach drift are believed responsible for areal differences in mineral composition. Distribution of certain selected minerals indicates that major sand movement trends are convergent toward the center section of this stretch of beaches.

## INTRODUCTION

The shoreline along the southwest coast of Rhode Island is marked by some 24 miles of continuous beach which is exposed to the open sea (fig. 1). At the westernmost end of this shoreline a former island, Napatree Point, is tied to the mainland by a sand bar. From Watch Hill Point to Point Judith this coastline is characterized by a series of sand bars which join a number of headland projections. Because of the irregularity of the mainland, these bars have produced several salt and brackish water ponds and marsh areas along the coast. At the present time only the larger ponds have normal exchange with the open sea and in each instance it is necessary to maintain these inlets by protective stone jetties or periodic dredging. Just west of Point Judith a perimeter of breakwaters shield approximately 1.5 miles of beach.

The beaches of southwestern Rhode Island contain a wide range of particle sizes. In general these sizes vary from cobbles to fine sand but at some locations boulders are also present. The Beach Erosion Board reports similar textures for these beaches (U. S. Congress, 1950, p. 11). Apparently the beach slopes are rather constant along the shoreline showing values which range randomly between 6° and 11° with most readings lying between 8° and 10°. Although there are some local variations in offshore slope seaward of the beach, the general bottom inclination immediately adjacent to the shore is not greatly different from one end of the shoreline to the other.

Behind the beaches and ponds a belt of hummocky terminal moraine composed of gravel, sand, silt, and clay extends across the coastal region of southern Rhode Island. This moraine ridge intersects the shore at Watch Hill Point but then diverges slightly from the shoreline for a distance of some 20 miles (fig. 1). From this area another segment of this moraine belt bends back to the shore at Point Judith (J. P. Schafer, personal communication, 1958). Between these segments of terminal moraine and the beaches, glacial deposits of ground moraine and outwash occur, and even granite outcrops are found at a number of localities (J. P. Schafer, personal communication, 1959). It is noteworthy that granite outcrops have been observed on the beach at both Weekapaug and Quonochontaug Points (Nichols and Marston, 1939, p. 1362; J. P. Schafer, personal communication, 1959). The low hills of ground moraine which form the headlands at Weekapaug, Quonochontaug, and Green Hill are composed of heterogeneous mixtures of gravel, sand, silt, and clay. At Matunuck Point a gravel and sand till overlies a stratified deposit of sand and silt. At the present time, under normal climatic conditions, only the headlands at Matunuck Point are producing any source material for the beaches.

The purpose of this study was twofold:

[1] Contribution No. 31 from the Narragansett Marine Laboratory. Manuscript received September 8, 1959.

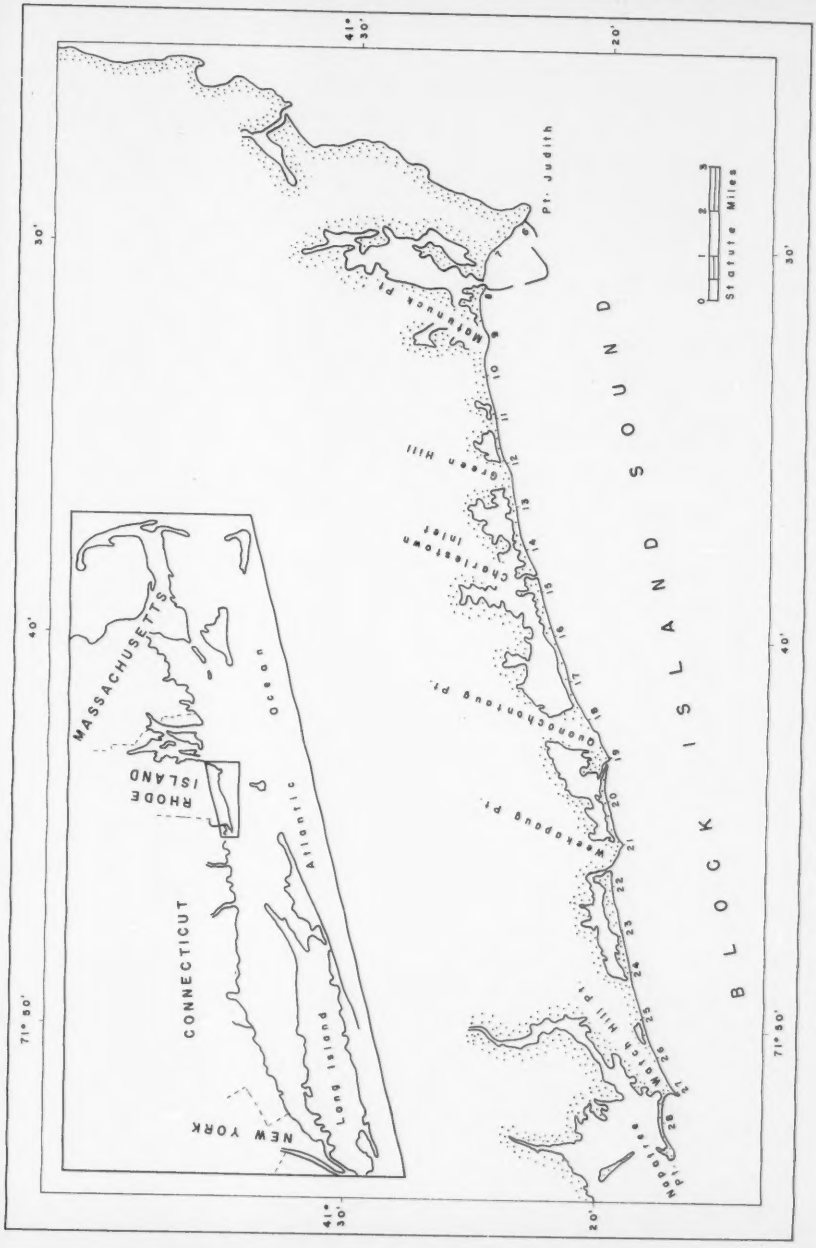


FIG. 1.—Outline map of the southwest coast of Rhode Island with beach sample locations.

(1) to determine the general composition of southwestern Rhode Island beach sands and (2) to ascertain whether any areal compositional differences can be used to define the pattern of local beach drift.

## FIELD METHODS

### Sampling of Beach Sands

The beach sand samples were collected at one-mile intervals from Point Judith to Napatree Beach west of Watch Hill (fig. 1). Samples 8–28 were taken during a three-day period in June 1955 when the sea and wind conditions were rather constant (U. S. Department of Commerce, Weather Bureau, 1955; S. D. Hicks, personal communication, 1959).

At each one-mile sampling location, a total of four individual small samples were collected from Bascom's mid-tide reference zone (1951, p. 867) and combined into one composite to represent each station. Pairs of these small samples were taken on two lines 3 ft apart. At each of these sites, the surface sand grains were brushed off and the uppermost sedimentation unit was sampled.

## LABORATORY INVESTIGATION

### Grain Size Determination

The beach samples were analyzed for particle size by sieving. Each sample was placed in a nest of sieves and shaken for 10 minutes. Quartile measures were used in describing these results. Sorting was calculated from Trask's (1932, p. 71) coefficient.

### Mineralogical Analysis

The general preparation of samples for mineral studies was as follows: Bulk samples of 10–20 gm were separated into heavy and light mineral fractions with bromoform (sp. gr. 2.85). Small amounts of magnetite were detected and removed from each heavy mineral fraction by means of a horseshoe magnet. The remaining heavy minerals were mounted in arochlor ($n = 1.66$), and these mineral species were identified under the petrographic microscope by optical properties and descriptions of their appearance.

A selective staining technique was utilized to separate quartz, potash feldspar, and plagioclase feldspar. This procedure was based in part on the method developed by Keller and Ting (1950, p. 124) and was previously reported (McMaster, 1954, p. 47). Each light mineral fraction was split to approximately 2 gm by means of an Otto microsplitter. This fraction was placed in a lead dish and bathed in warm concentrated hydrofluoric acid for one minute. After thorough washing the material was immersed in a 1 percent aqueous solution of malachite green for 5 minutes, rinsed, immersed in a saturated solution of sodium cobaltinitrite for 5 minutes, rinsed, and dried. The material was again split and mounted in Canada balsam. Quartz remains unchanged, potash feldspar stains yellowish, and plagioclase stains green.

The relative proportions of the different detrital minerals were determined by counting carefully spaced fields of grains arranged systematically over the area of the slide. A minimum of 300 grains were tabulated for each count, and in most samples 400 grains were counted.

## MINERALOGY AND GRAIN SIZE

### Minerals Present

A large number of minerals is found in southern Rhode Island beach sands. This assemblage is presented in table 1. However, this list of minerals is probably incomplete because only a small number of grains were examined from each bulk sample. The distribution of several selected minerals along the beach is shown in figure 2 and table 2.

In compiling mineral frequencies certain varieties were grouped to simplify tabulation. Actinolite, hornblende, and tremolite

TABLE 1.—*Minerals in the beach sands*

| | |
|---|---|
| Actinolite | Kyanite |
| Andalusite | Leucoxene |
| Apatite | Limonite |
| Augite | Magnetite |
| Biotite | Monazite |
| Chlorite | Muscovite |
| Chloritoid | Quartz |
| Clinozoisite | Rutile |
| Corundum | Sillimanite |
| Diopside | Staurolite |
| Epidote | Titanite |
| Potash Feldspar | Tourmaline |
| Soda-Lime Feldspar | Tremolite |
| Garnet | Xenotime |
| Hornblende | Zoisite |
| Hypersthene | Zircon |

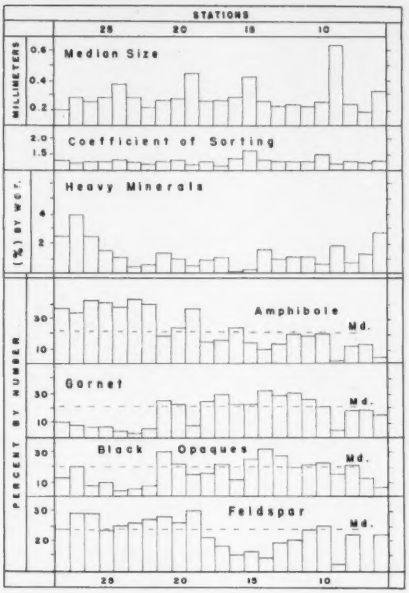


FIG. 2.—Distribution of selected minerals and certain size parameters along southern Rhode Island beaches.

are grouped under the heading of amphibole, whereas black opaques include all black opaque minerals with the exception of magnetite which had been previously removed.

### Particle Size

The general results of the mechanical analysis are presented in figure 2. From this plot, it is obvious that most of the median size values lie between 0.2 mm and 0.3 mm, and there is no apparent trend in the areal distribution of these values along the shore. In addition, sorting also appears to be quite constant over this stretch of beach. It is believed that these findings suggest a uniformity of certain environmental conditions in this coastal region.

## STATISTICAL ANALYSES OF COMPOSITION

### Multivariate Analysis

With these mineral analyses it is necessary to decide whether significant differences occur among samples based on the evaluation of all or several of the listed constituents. Recently Krumbein and Tukey (1956) proposed a multivariate method of statistical analysis which permits the simultaneous treatment of several variables for problems involving areal differences in composition.

In the present study, the writer was interested in learning whether there is a greater variation in mineral composition between certain areas along the shore than there is between localities within these areas and whether there is more variation between these localities than there is between individual samples within these localities.

The following procedure was pursued: Based on the plot of percent by number of mineral species against distance (fig. 2), three major beach segments of possible different composition (A, B, and C) were suggested in the stretch of shore between the breakwater near Matunuck Point and Napatree Point (fig. 1). Within each of these segments the following localities were chosen: For A, stations 8 and 10; for B, stations 14 and 15; and for C, stations 26 and 27. By making a completely different mineral analysis for each of these stations, a total of two counts was available for each locality. The mineral species selected and their abundance are shown in tables 2 and 3.

In the manner described by Krumbein and Tukey (1956, p. 334), these mineral percentage data were normalized by the arc sine square root transformation. After these transformed values were condensed and mean squares computed, the results were introduced into the multivariate model (table 4). Then the interaction mean squares were determined and significant tests were set up by taking ratios of these mean squares to obtain F values. These were tested against critical values of F for the indicated degrees of freedom at the 5 percent level of significance (Dixon and Massey, 1957, p. 390), and appropriate statistical hypotheses were accepted or rejected.

Tables 4 and 5 summarize the results of the variance analysis for the selected heavy and light mineral fractions in segments A, B, and C. From these data it is inferred at the 5 percent significance level that (1) the variation in composition between segments is significantly greater than the variation in the composition between localities

TABLE 2.—*Percentages of heavy minerals in Rhode Island beach sands*

| Segment | A | | | | B | | | | C | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Locality | 1 | | 2 | | 1 | | 2 | | 1 | | 2 | |
| Count | a | b | a | b | a | b | a | b | a | b | a | b |
| Amphibole | 12 | 14 | 21 | 18 | 10 | 12 | 15 | 17 | 44 | 44 | 34 | 35 |
| Biotite | 5 | 2 | 4 | 1 | — | — | 5 | — | 1 | — | — | — |
| Chlorite | 18 | 10 | 13 | 5 | 3 | 1 | 10 | 11 | 6 | 2 | 4 | 2 |
| Epidote | 1 | 2 | — | 2 | 2 | 2 | 2 | 1 | 2 | 2 | 2 | 5 |
| Garnet | 18 | 26 | 21 | 25 | 33 | 29 | 22 | 30 | 7 | 6 | 8 | 9 |
| Staurolite | 5 | 8 | 5 | 4 | 4 | 8 | 5 | 6 | 3 | 2 | 2 | — |
| Titanite | — | 2 | 1 | 2 | 4 | 2 | 1 | 3 | 1 | — | 1 | — |
| Zircon | — | — | 3 | 2 | 3 | 3 | 3 | 3 | — | 1 | 2 | 1 |
| Blk Opaques | 21 | 18 | 22 | 25 | 32 | 34 | 25 | 18 | 8 | 9 | 21 | 19 |
| Others | 20 | 18 | 10 | 16 | 9 | 9 | 12 | 11 | 28 | 34 | 26 | 29 |
| Total | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 | 100 |

TABLE 3.—*Percentages of light minerals in Rhode Island beach sands*

| Segment | A | | | | B | | | | C | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Locality | 1 | | 2 | | 1 | | 2 | | 1 | | 2 | |
| Count | a | b | a | b | a | b | a | b | a | b | a | b |
| Quartz | 76 | 74 | 75 | 75 | 85 | 81 | 83 | 83 | 70 | 71 | 71 | 70 |
| Feldspar | 22 | 24 | 25 | 25 | 14 | 18 | 16 | 17 | 29 | 29 | 29 | 30 |
| Others | 2 | 2 | — | — | 1 | 1 | 1 | — | 1 | — | — | — |

TABLE 4.—*Rhode Island beach sand (segments A, B, and C; heavy minerals)*
*Analysis of variance*

| Source | Sum of Squares | Degrees Freedom | Mean Square | F |
|---|---|---|---|---|
| Between segments | 68.41 | 2 | | |
| Between localities within segments | 11.08 | 3 | | |
| Between counts within localities | 18.13 | 6 | | |
| Between types | 10,208.39 | 9 | | |
| S X T | 2,437.50 | 18 | 135.41 | 5.85 (s)* |
| L X T | 624.73 | 27 | 23.13 | 2.51 (s)* |
| C X T | 495.87 | 54 | 9.18 | |
| Total | 13,864.11 | 119 | | |

(Computed from arc sine transformation.)
* s denotes significance at 5% level.

in a segment and (2) the variation in composition between localities is significantly greater than the variation in composition between counts.

When one appraises these inferences it becomes evident that although important variations in composition are implied, there is no indication as to where these differences occur along this shore because the three segments were treated collectively. Therefore in order to define the area locations of compositional differences, multivariate analyses were repeated utilizing separate two-segment tests, A–B, A–C, and B–C. These

TABLE 5.—*Rhode Island beach sand (segments A, B, and C; light minerals) Analysis of variance*

| Source | Sum of Squares | Degrees Freedom | Mean Square | F |
|---|---|---|---|---|
| Between segments | 4.56 | 2 | | |
| Between localities within segments | 19.52 | 3 | | |
| Between counts within localities | 8.51 | 6 | | |
| Between types | 20,049.27 | 2 | | |
| S X T | 331.85 | 4 | 82.96 | 7.73 (s)* |
| L X T | 64.42 | 6 | 10.73 | 3.56 (s)* |
| C X T | 35.90 | 12 | 2.99 | |
| Total | 20,514.03 | 35 | | |

(Computed from arc sine transformation.)
* s denotes significance at 5% level.

results show that for the heavy mineral analysis of variance, segment C is significantly different from either A or B. However, the variation in composition between segments A–B and localities is not significant although the variation between localities and counts is significant at the 5 percent significance level (tables 6 and 7).

From these more exhaustive tests of the mineral data, the evidence suggests that within the stretch of beach between the breakwater near Matunuck Point and Napatree Point only two areas of beach are significantly different.

## Trend Analysis

A nonparametric technique of statistical analysis, a runs test (Hald, 1952, p. 342), was used to test the hypothesis of random distribution of certain beach minerals. This approach was utilized to demonstrate whether the plots of percentages of selected mineral species against distance along the shore characterize random distribution or reflect definite trends along the beach. Accordingly the amounts of amphibole, garnet, feldspar, and black opaques (fig. 2) were listed for the sequence of samples between the breakwater (station 8) and Napatree Beach (station 28). Next a median was calculated for each set of percentages, runs determined above and below the median, and length of runs tabulated in the manner outlined by Hald (1952, p. 351) (fig. 2 and table 8).

Randomness was tested by using the length of run. The formula,

$$k_{.05} \simeq 3.3(\log n + 1) \text{ (Hald, 1952, p. 347)}$$

where $k$ equals the expected length of run at the 5 percent significance level and $n$ equals the number of observations, provided

TABLE 6.—*Rhode Island beach sand (segments A and B, heavy minerals) Analysis of variance*

| Source | Sums of Squares | Degrees Freedom | Mean Squares | F |
|---|---|---|---|---|
| Between segments | 0.04 | 1 | | |
| Between localities within segments | 10.88 | 2 | | |
| Between counts within localities | 5.17 | 4 | | |
| Between types | 6,368.14 | 9 | | |
| S X T | 356.39 | 9 | 39.59 | 1.63 (ns)* |
| L X T | 435.95 | 18 | 24.21 | 2.22 (s)* |
| C X T | 392.71 | 36 | 10.90 | |
| Total | 7,569.28 | 79 | | |

(Computed from arc sine transformation.)
* s denotes significance, ns nonsignificance at 5% level.

TABLE 7.—*Rhode Island beach sand (segments A and B, light minerals) Analysis of variance*

| Source | Sum of Squares | Degrees Freedom | Mean Square | F |
|---|---|---|---|---|
| Between segments | 0.06 | 1 | | |
| Between localities within segments | 17.36 | 2 | | |
| Between counts within localities | 4.17 | 4 | | |
| Between types | 14,023.23 | 2 | | |
| S X T | 126.05 | 2 | 63.02 | 4.31 (ns)* |
| L X T | 58.37 | 4 | 14.59 | 4.97 (s)* |
| C X T | 23.45 | 8 | 2.93 | |
| Total | 14,252.69 | 23 | | |

(Computed from arc sine transformation.)
* s denotes significance; ns nonsignificance at 5% level.

TABLE 8.—*Distribution of runs below and above the median according to length*

| Length of Run | Number of Runs | | | |
|---|---|---|---|---|
| | Amphibole | Black Opaques | Feldspar | Garnet |
| | Below (Above) Median | Below (Above) Median | Below (Above) Median | Below (Above) Median |
| 1 | 2 | 6 | 2 | 1 |
| 2 | 2 | 2 | 1 | 2 |
| 3 | — | — | 1 | — |
| 4 | — | — | — | — |
| 5 | — | 1 | — | — |
| 6 | — | 1 | 1 | — |
| 7 | 1 | — | — | 1 |
| 8 | 1 | — | 1 | 1 |
| Total | 6 | 10 | 6 | 5 |
| No. of observations | 21 | 21 | 21 | 21 |

the necessary statistic. By substituting the total number of observations ($n$) for each selected mineral (table 8) in this formula, the greatest length of run ($k$) was determined for each situation (table 9). These values defined the critical lengths within expected random distribution for the given number of stations at the specified level of significance. By comparing these values with the measured length (table 9) the hypothesis of randomness was accepted or rejected.

TABLE 9.—*Analysis of length of run*

| Mineral | Expected length of run ($K_{.05}$) | Observed length of run | Significance |
|---|---|---|---|
| Amphibole | 7.7 | 8 | s. |
| Black Opaques | 7.7 | 6 | n.s. |
| Feldspar | 7.7 | 8 | s. |
| Garnet | 7.7 | 8 | s. |

From these data, it is inferred that a run of length 8 is significant and the distributions for amphibole, feldspar, and garnet with these lengths show definite trends along the beach.

## DISCUSSION

When the general regularity of the grain size data (fig. 2) from the uppermost beach sediment unit is examined together with the oceanographic factors and field sampling techniques, there is a strong suspicion that the beach environmental conditions were essentially uniform over this 20-mile stretch of shore at the time of sampling. Otherwise one might expect more marked areal changes

in particle size due to variations in waves and currents, beach profile, or different sources of supply along the shoreline.

However, the study of mineral composition of these same sands and statistical treatment of their mineralogical data clearly demonstrate that (1) this stretch of beach can be separated into two individual segments which are mineralogically different and (2) certain minerals (amphiboles, feldspars, and garnet) are not randomly distributed along the shore but show definite trends. Thus it is evident that these variations in mineral composition are unrelated to differences in sample grain size and therefore are real and significant.

A preliminary study of the heavy mineral composition of several glacial deposits in coastal southern Rhode Island reveals that there is a marked change in the nature of these minerals at Watch Hill when compared to those at Matunuck Point and Point Judith. In the Watch Hill area the glacial material is characterized by heavy minerals which contain a high proportion of amphiboles and low percentages of garnet and black opaques. Toward the east, these glacial samples show a decrease in amphiboles and corresponding increases in black opaques and garnet. At Matunuck Point and Point Judith the amount of amphiboles in the heavy fractions is low. Therefore the character of source material in this region accounts for the gross differences in mineral composition which appear along this stretch of shoreline and is an important consideration in the problem of beach sand movement

Some information on the pattern of littoral drift along this coastline is available (U. S. Congress, 1950, p. 19). The Beach Erosion Board reports that prior to the construction of the Point Judith breakwaters littoral drift moved westward from Point Judith to Matunuck Point. As sand has been accumulating on the west side of the westernmost breakwater since this construction, there has been a local reversal to an eastward drift between Matunuck Point and the breakwater. For the stretch of shore from Matunuck Point to Napatree Point the Beach Erosion Board believes that there is no conclusive evidence of a predominant direction of beach drift.

In addition to the Point Judith breakwater, other structures such as jetties and seawalls have been erected along this shoreline and all these structures have had varying effect on the natural processes along the beach. The pattern of drift to be outlined here is an outgrowth of these modified shoreline processes. However, no attempt will be made to include the beach protected by the breakwaters near Point Judith.

Based on the distribution of beach samples and their mineralogical data, it is believed that the pattern of sand movement along the southern Rhode Island shore takes the following general form: movement is convergent toward the Charlestown Inlet area and divergent at least from the vicinity of Matunuck Point (fig. 3). No effort will be made to evaluate the strengths of these driftings.

At the eastern end of this stretch of beach, it has been established that sand moves from Matunuck Point toward the Point Judith breakwater in an easterly direction (U. S. Congress, 1950, p. 19). The writer believes that a nodal zone exists in the general vicinity of station 10 (fig. 1), a mile or so west of Matunuck Point. From this zone sand moves eastward toward the westernmost breakwater near Point Judith and westward in the direction of the Charlestown Inlet (fig. 3). This inference is based on (1) the statistical test of lengths of run for garnet and feldspar and (2) the systematic variations in mineral percentages when plotted against distance as shown in figure 2. Although this statistical test in no way implies a normal progression in mineral abundance, the fact that a significant trend originates in this area seems highly suggestive. Because there are only two stations between this nodal zone and the breakwater, one would hardly expect pertinent information on mineral trends toward the east. It is noteworthy that the source sample collected on the backshore just east of Matunuck Point contains a heavy mineral assemblage very similar to that of station 8.

From the nodal zone, the littoral transport can be traced westward to the Charlestown Inlet area (fig. 3). Here it is believed that sand movement may be oscillatory, showing neither an eastward nor westward trend. It seems that the general low amount of the heavy mineral group as well as the

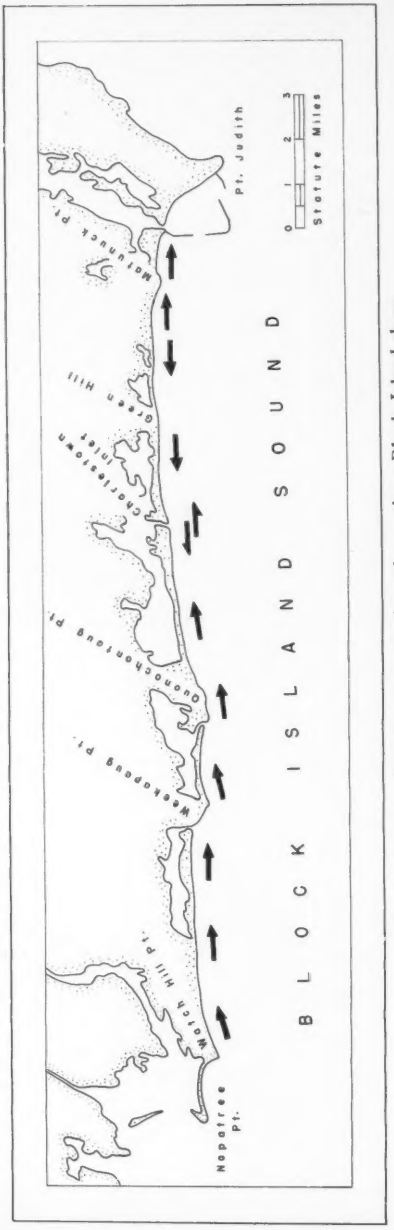


FIG. 3.—Pattern of sand movement along the southern Rhode Island shore.

low abundance of amphiboles and feldspars and higher proportions of black opaques and garnet indicate this probability.

In addition, the central location of the Charlestown Inlet in reference to the adjacent headlands also would seem to support this contention. With the exception of the inlet at Charlestown there are no natural or artificially maintained breachways to sap this westward drift.

Whether or not the Watch Hill Point area acts as another nodal zone is still an open question. Certainly one station between Napatree and Watch Hill Points is not sufficient to demonstrate a trend. However, the decrease in heavy mineral percentages eastward from Watch Hill Point with a general uniformity of grain size as well as over-all decrease in abundance of amphiboles and feldspars and increase in garnet and black opaques seem to reflect an eastward drift away from the source area. Further, it is believed that this predominant eastward drift continues into the Charlestown Inlet area. With the construction of jetties at Weekapaug Breachway, sand is now accumulating on the western side of the inlet, and this is certainly conclusive evidence of an eastward littoral transport. However, the presence of this obstruction also may be responsible for the continuous high percentages of amphiboles and low amounts of garnet and black opaques between stations 22 and 27 and the rather abrupt changes in the proportion of these minerals toward the east (fig. 3). In addition, it is well to note that the inlets at both Weekapaug and Quonochontaug are found adjacent to their eastern bordering headlands (fig. 3), and these locations, too, strongly suggest the occurrence of a predominantly eastward drift along this stretch of shoreline. In the past both these inlets probably sapped a rather large proportion of the beach drift but at present only the inlet at Quonochontaug is maintained by natural processes.

## CONCLUSIONS

The following conclusions can be drawn from the present study:

1.—A great variety of minerals occur in southern Rhode Island beach sands. It is recognized that the list presented in this paper is undoubtedly incomplete. The most

abundant mineral species or mineral groups are amphibole, chlorite, garnet, staurolite, and black opaques in the heavy fraction and feldspar and quartz in the light fraction.

2.—Areal differences in mineral composition along the shoreline are attributed to differences in the character of the source materials and to the pattern of beach drift.

3.—The areal distribution of certain selected minerals indicates that sand movement along the southern Rhode Island shore takes the following general pattern: sand is transported eastward from the Watch Hill Point area and westward from a nodal zone near Matunuck Point, thus converging in the Charlestown Inlet area. It is believed that in the area around the breachway at Charlestown sand movement is oscillatory with no predominant trend.

ACKNOWLEDGMENTS

The writer is particularly indebted to David E. Frazier, formerly of the Narragansett Marine Laboratory, for his aid in collecting the beach samples and also for preparing these field samples for mineralogical analysis. J. S. Kahn's and Saul B. Saila's assistance with the statistical analyses is gratefully acknowledged.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 414–431
FIGS. 1–11, SEPTEMBER, 1960

# PLEISTOCENE ALGAL PINNACLES AT SEARLES LAKE, CALIFORNIA[1]

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## ABSTRACT

Searles Lake is a playa or dry lake which occupies the central floor of Searles Basin, a northward-trending graben of the Basin and Range province in southeastern California. During pluvial periods of the Pleistocene epoch deep-water lakes filled Searles Basin, and in two of these lakes more than 500 pinnacled masses of calcareous tufa accumulated in an arm-like bay at the southwest end of the basin.

Pinnacles are tower-, tombstone-, and cone-shaped masses and large limestone ridges. Most of the pinnacles are 10 to 40 ft high, but a few reach heights between 100 and 140 ft. Basal diameters or widths range from about 10 ft to as much as 500 ft and average 20 to 30 ft. Seven varieties of tufa compose the pinnacles; one of these varieties is also found in lenticular bodies buried in the lacustrine sediments underlying the pinnacles at the northern end of the bay.

It is proposed that the pinnacles were precipitated by algae about the orifices of sublacustrine springs issuing along faults striking N. 65° W., N. 50° W., N. 30° E., N. 55° E., and N. 65° E. in the underlying basement rocks. Deposition of the pinnacles at the southwest end of the bay took place in a lake of Tahoe age which filled Searles Basin more than 32,000 years ago. Pinnacles midway along and at the northern end of the bay formed in a lake of Tioga age which initially flooded Searles Basin about 23,000 years ago and lasted until about 10,000 years ago. Tioga Searles Lake at its maximum stand was approximately 460 ft deep and at its highest stage the surface reached an elevation of about 2000 ft.

## INTRODUCTION

Spectacular pinnacles of calcareous tufa rise above gently dipping lacustrine sediments at Searles Lake in southeastern California. At the southwest end of this desiccated Pleistocene lake more than 500 of these limestone spires are concentrated in an area of about 16 square miles.

Brief references to the Searles Lake pinnacles have appeared periodically (Free, 1914; Gale, 1915; Blackwelder, 1941 and 1954), but no thorough description has yet been presented. As a consequence, the geologic significance of the Searles pinnacles has previously received little attention, which is somewhat surprising considering that studies of pluvial tufas have proven fruitful in reconstructing the past hydrology of other Pleistocene lake basins (Russell, 1885; Broecker and Orr, 1958). The present study presents a detailed description of the pinnacles at the southwest end of Searles Basin, and from their postulated origin and depositional history general conclusions are drawn regarding the late Pleistocene hydrology of Searles Lake.

### Location and Physiography of Searles Lake

Searles Lake is a playa or dry lake of the moist crystal body type (Stone, 1958) which lies approximately 125 miles northeast of Los Angeles. The playa occupies the central floor of Searles Basin, a 25-mile long and 10-mile wide northward-trending graben of the Basin and Range province. Although there is presently no perennial lake in Searles Basin, during the Pleistocene epoch rains of more pluvial times and glacial melt water from the eastern Sierra Nevada periodically formed large deep-water lakes in the basin (Smith and Pratt, 1957). These late Pleistocene lakes and drainage into the basin have been discussed by Gale (1915), Blackwelder (1941 and 1954), and more recently by Smith and Pratt (1957), Flint and Gale (1958), and Smith (1958).

## DESCRIPTION OF THE PINNACLES

### Distribution

Pinnacles of calcareous tufa occur in a number of areas within Searles Basin and in adjacent Salt Wells Valley to the west. However, only the pinnacles that are con-

[1] Publication authorized by the Director, U. S. Geological Survey. Manuscript received November 6, 1959.

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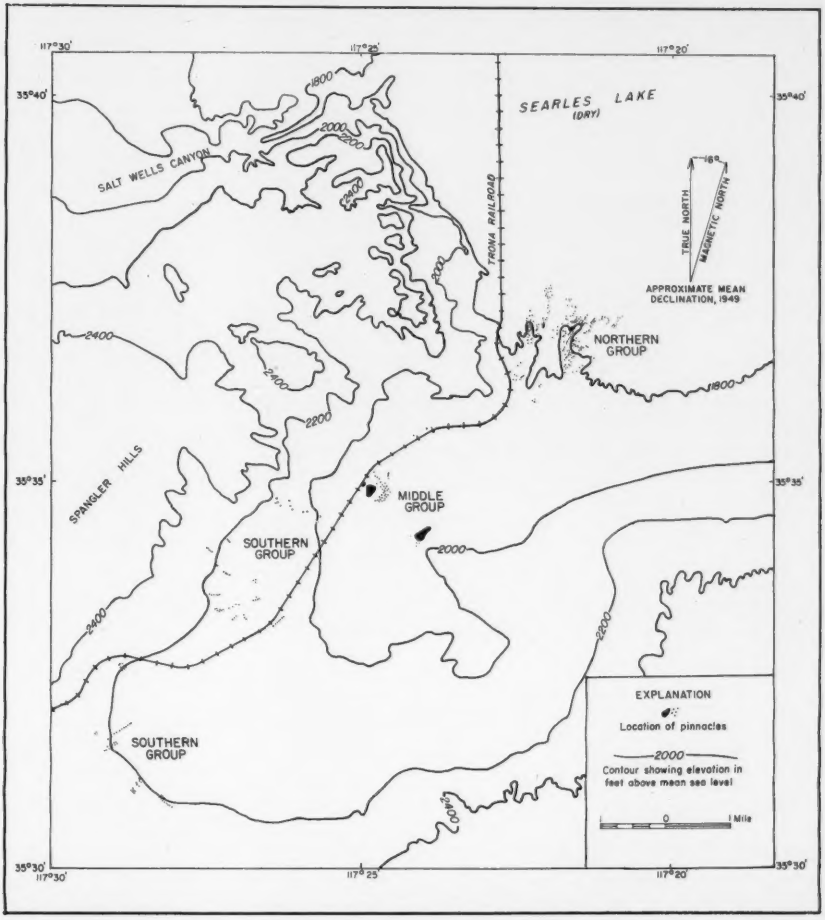


FIG. 1.—Sketch map showing distribution of most of the pinnacles at the southwest end of Searles Basin; however, less than half of the pinnacles in the southern group are shown. The two enlarged black areas in the middle group are the large limestone ridges; the northern limestone ridge is also shown.

centrated at the southwest end of Searles Basin are described and discussed below.

Pinnacles at the southwest end of the basin are within a 10-mile long and 4-mile wide southwestward-trending arm-like bay of the old Pleistocene lakes (fig. 1). They are confined largely to a northern group at the head of the bay, a middle group about midway along the length of the bay, and a southern group throughout most of the southwest part of the bay. The distribution of most of the pinnacles as determined from aerial photographs is shown on figure 1; however, less than half of the southern pinnacles are shown because most of these are too small to show on the photographs.

Pinnacles of the northern group are restricted to a roughly circular area of about 2 square miles which includes approximately 200 pinnacles. The bases of the northern pinnacles are at elevations between 1720 and 1850 ft; the lowest point on the present desiccated lake floor is at an elevation of 1616 ft. Pinnacles of the middle group are also within an area of about 2 square miles and are about 100 in number. Elevations at the bases of these pinnacles range from about 1920 ft to approximately 2000 ft. The southern group comprises approximately 200 pinnacles within an area of about 12 square miles. These spires are at elevations between about 2100 ft and 2260 ft.

### Alignments

Pinnacles are arranged along lines that trend roughly N. 65° W., N. 50° W., N. 30° E., N. 55° E., and N. 65° E. These trends can be discerned in the field, but they are best shown on aerial photographs. The N. 65° W. and N. 50° W. lineations are developed only among the southern pinnacles and the N. 65° E. lineation only in the northern group; the N. 30° E. and N. 55° E. alignments are the strongest trends in all groups.

### Shapes and Dimensions

The pinnacles can be classified into four general shapes: tower-shaped spires with roughly circular horizontal cross sections, tombstone-shaped structures with ellipsoidal cross sections, small cone-shaped masses and large limestone ridges (figs. 2, 8, 9, 10, and 11).

Tower and tombstone forms are most common in the northern pinnacles and are about equal in abundance (table 1). Cone-shaped pinnacles are slightly less numerous and are most abundant along the northern (lakeward) periphery of the northern group (fig. 11). Among the middle and southern groups tower- and cone-shaped pinnacles are most common. Tombstone-shaped masses are absent, but elongated low mounds somewhat similar to the northern tombstone structures occur rarely in the middle group. Many of the tower-shaped pinnacles of the southern group are closely spaced and form linear castellated ridges. There are only three large limestone ridges. One of these is in the northern group and the two in the middle group are the largest pinnacles in Searles Basin (figs. 1 and 10).

Heights of the pinnacles range from about 1 ft to 140 ft. The average height of the northern pinnacles is 30 to 40 ft; however, pinnacles of the middle and southern groups are usually betwen 5 and 15 ft tall, although the highest pinnacle (140 ft) is in the middle group. Cone-shaped pinnacles are the smallest and commonly are less than 10 ft high.

Basal diameters or widths range from about 10 ft to as much as 500 ft, with an average of 20 to 30 ft. Many tombstone pinnacles are 60 to 70 ft long and 20 to 35 ft wide. The limestone ridge of the northern group is approximately 500 ft long, 100 ft wide, and 120 ft high (figs. 1 and 2).

The sides of most of the pinnacles are nearly vertical, but they gradually taper upward to rounded, flat, or castellated summits. The lowest third of many of the northern pinnacles is hidden beneath flanking basal slopes of lacustrine sediment which extend outward and downward from the pinnacles at angles of 25 to 30 degrees. Radially outward from the pinnacles the flanking slopes dip less steeply and gradually merge with the nearly horizontal lake sediments surrounding and underlying the pinnacles (figs. 2 and 7). Flanking basal slopes are not well developed or are absent in the middle and southern groups.

### Types of Tufa Forming the Pinnacles

Seven varieties of tufa form the Searles pinnacles (fig. 3). These are: (1) a porous and somewhat cavernous cream-colored *stony lithoid* tufa, (2) a highly porous cream-colored *cavernous lithoid* tufa, (3) a light-gray somewhat porous compact *massive lithoid* tufa, (4) a rather dense branching or arborescent *dendritic* tufa, (5) a cream-colored *nodose* tufa, and (6) a chalk-white to cream-colored *tubular* tufa which grades upward into (7) banded *lobate* tufa.

*Stony, cavernous, and massive lithoid tufa.* —Russell's term, *lithoid tufa* (Russell, 1887, p. 311), which was applied to the stony-appearing tufas at Lahonton and Mono

TABLE 1.—*Character of Searles Lake pinnacles*

| Shape | Dimensions | Structure and composition | Abundance and distribution |
|---|---|---|---|
| | | Northern group | |
| Tower | Roughly circular cross sections; usually twice as tall as wide. Average pinnacle is 40 ft high and 20 ft in diameter. | Most consist of stony lithoid tufa; some have an outer cover of cavernous lithoid tufa from 1 to 20 ft thick. Bases of pinnacles are encircled by 2 to 5 ft of nodose and tubular tufa. Pinnacles near the southern margin consist of cavernous tufa surrounding columns of tubular tufa. | Most numerous in eastern half of pinnacle area; they constitute approximately 35 to 40 percent of northern pinnacles. |
| Tombstone | Ellipsoidal in cross section. Usually three times as long as wide and twice as tall as wide. Typical pinnacle is 40 ft high, 60 ft long, and 20 ft wide. | All are composed of stony lithoid tufa, some have remnants of an outer cavernous lithoid layer. Base of pinnacle encircled by 2 to 5 ft of tubular and nodose tufa. | Most numerous in western half of pinnacle area. Usually occur in linear groups. They form about 30 percent of northern pinnacles. |
| Cone | Have roughly circular basal cross sections. Usually 10 to 15 ft tall and approximately 30 ft in diameter. | All have central cores of very porous lithoid tufa which range from 2 to 6 ft in diameter. Surrounding the core are vertical layers of nodose tufa totaling 20 to 35 ft in thickness. | Most common near the northern or lakeward margin of the pinnacle area. They comprise about 30 percent of northern pinnacles. |
| Ridge | Very large feature, several hundred feet long, about 100 ft wide and 100 ft high. | Consists of stony lithoid tufa with some vestiges of an outer cavernous lithoid layer. Summits of the ridge are castellated by vertical pillars of tubular tufa from 5 to 15 ft high. | Rare, only one present among northern group. |
| | | Middle group | |
| Tower | Columnar in shape, usually have flat or rounded summits. Commonly 20 to 25 ft high and 15 to 20 ft in diameter. | Consist of banded cavernous lithoid tufa surrounding and overlying vertical columns of tubular tufa a few inches to 6 ft in diameter. | Form about 40 percent of the pinnacles. Most are grouped about the largest of the two limestone ridges. |
| Cone | Small mounds 2 to 15 ft high and 20 to 30 ft in diameter. A few are elongated and have ellipsoidal cross sections. | Consist of cavernous lithoid tufa. Some are remains of collapsed tower-shaped pinnacles. | Include approximately 50 to 55 percent of the pinnacles. They occur throughout the pinnacle area. |
| Ridge | Very large features; several hundred to more than 1000 ft long, 400 to 500 ft wide and 100 to 140 ft high. | Constructed of porous hummocky outward-dipping layers of massive lithoid tufa. Cracks and joints are filled with tubular and digitate tufa and the summit and sides of the ridges are coated with swirling masses of tufa. | Although very conspicuous features, only two occur. Smaller outlying mounds of the massive lithoid tufa are associated with the ridges. |

(*Continued on next page*)

TABLE 1.—(*Continued*)

| Shape | Dimensions | Structure and composition | Abundance and distribution |
|---|---|---|---|
| | | Southern group | |
| Tower | Irregular castle-like masses 5 to 25 ft high, 5 to 40 ft in diameter. | Constructed of deeply weathered cavernous lithoid tufa usually showing horizontal banding. | Forms about 50 percent of the pinnacles, many, however, have been destroyed by blasting and now appear as cone-shaped masses, most pinnacles occur in close-spaced linear groups. |
| Cone | Typically small masses 1 to 5 ft high, 10 to 20 ft in diameter. | Consist of deeply weathered lithoid tufa. | Constitute about 50 percent of the pinnacles, but possibly much more as these pinnacles are not easily seen in the field and are scattered throughout the entire area of the southern group. |

Lakes, has been adopted by the writer to refer to the stony, cavernous, and massive varieties at Searles Lake.

Both stony and cavernous lithoid tufas are banded (1 mm to 3 to 4 cm) and are constructed of a very porous and labyrinthine microcrystalline calcite (figs. 3b, 4, and 6); stony lithoid tufa, however, is the more massive and far less cavernous of the two. The bulk densities (gm/cc) of stony and cavernous lithoid tufa are 1.5 and 1.3, respectively.

Massive lithoid tufa is an exceptionally dense variety which forms the two large limestone ridges of the middle group and smaller outlying mounds closely associated

FIG. 2.—In the middle distance is a group of tombstone-shaped pinnacles towering above gentle northeast-dipping late Pleistocene lacustrine sediments. A large limestone ridge about 120 ft high is on the far right. View is to the west and includes approximately one-half of the northern pinnacles.

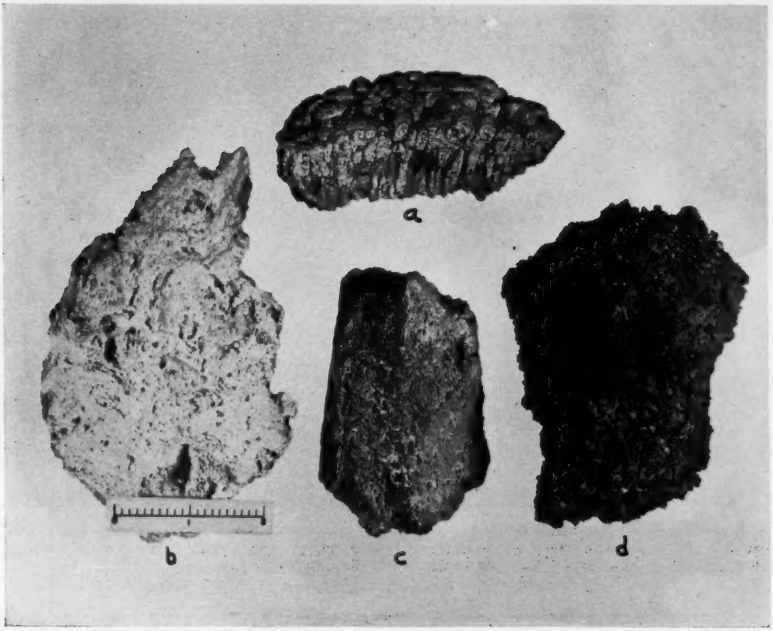


FIG. 3.—Types of tufa forming the Searles Lake pinnacles: (a) polished vertical section of tubular tufa grading upward into hemispherical lobate tufa; (b) stony lithoid tufa; (c) massive lithoid tufa from southern limestone ridges; and (d) nodose tufa. (Scale is 2 in long.)

with the ridges. Although this variety is highly compact, it is porous; and has a bulk density of 1.9 (fig. 3c). Most of the openings are 1 to 2 mm in diameter and are elongated normal to the outward-dipping bedding planes which form these ridges.

*Dendritic tufa.*—Russell's term, *dendritic tufa*, is also adopted to refer to the branching arborescent tufa which is typically associated with deposits of cavernous lithoid tufa. This variety consists of 5- to 10-mm thick bands of tightly packed columns (usually 1 mm in diameter) of upward-branching stems of maroon, green, cream, or gray tufa (fig. 5). Dendritic tufa is rather uncommon in the Searles pinnacles, but it usually forms draping festooned deposits near the bases of the pinnacles.

*Nodose tufa.*—Nodose tufa has a highly porous open-framework structure and consists of a profusion of small (0.5 to 5 mm) concentrically banded gnarls or nodes (fig. 3d). This variety is also layered, with most layers 4 to 8 inches thick.

*Tubular tufa.*—Tubular tufa (fig. 3a) occurs mainly in columnar bodies ranging in diameter from about 4 in to 6 ft (fig. 7). Tubules radiate outward and upward from the axes of the columns; the columns themselves are constructed of hemispherical layers or bands of the tubules from 1 to 3 in thick (fig. 9). Within the layers the tubules form small upward-branching groups; individual tubules range in diameter from about 3 to 10 mm, and expand or increase in diameter upward. The walls of the tubules are 0.5 to 1 mm thick and are bridged by thin (0.5 to 1 mm) upward-arched calcareous diaphrams separating small chambers 0.5 to 1.5 mm high. Many of these chambers are partly or completely filled with banded microcrystalline calcite which forms a porous digitate tufa. The shafts or passages which separate and lie parallel to the tubules

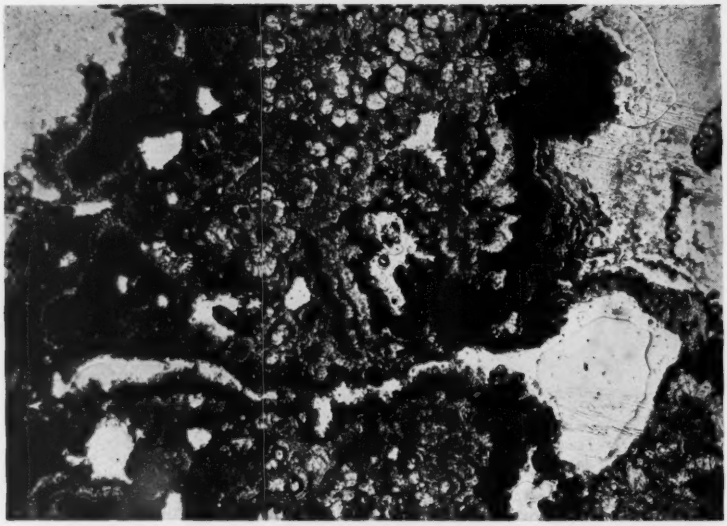

FIG. 4.—Cells and structures believed to be of algal origin (×38) in porous stony lithoid tufa. The channels and openings are lined with microcrystalline calcite rich in dark organic matter which suggests that filamentous algae may have grown actively along these passages.

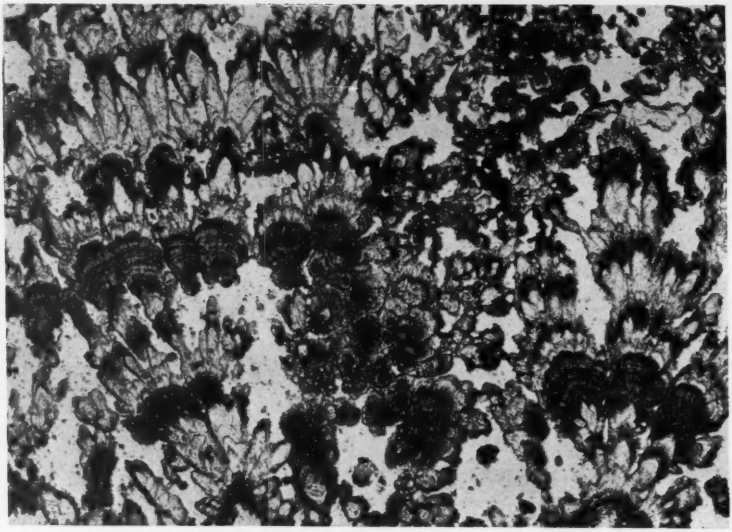

FIG. 5.—Photomicrograph (×24) of dendritic tufa. Dark-colored lobate layers of banded tufa are probably algae secreted, banded calcite laths radiating from these layers may be largely of inorganic origin. Conceivably, the layers represent annual accruements. The oval pigmented bodies near the center of the photograph may be algal cells.

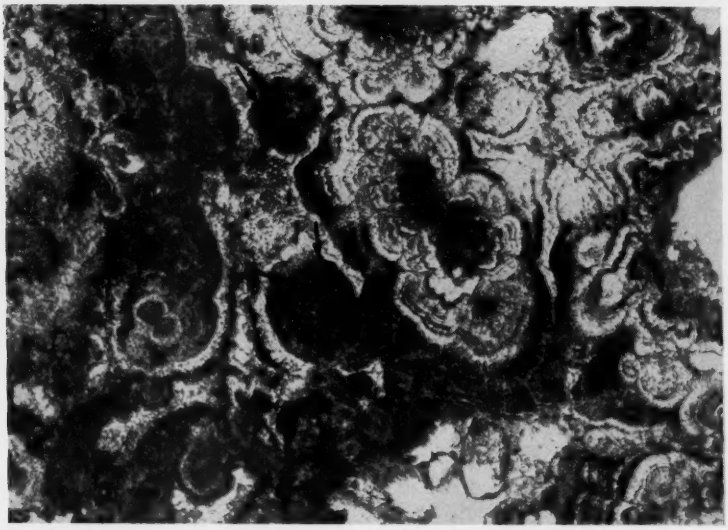

FIG. 6.—Photomicrograph (×50) of algal structures in stony lithoid tufa. The small highly pigmented oval bodies (arrows) are probably algal cells.

are filled with clastic mineral grains, oölites, relatively coarse-grained microcrystalline calcite, and rarely ostracod valves.

*Lobate tufa.*—Lobate tufa forms the outer ½ to 1 in of most of the tubular columns (fig. 9). This variety consists of banded hemispherical lobes ranging in diameter from 5 to 30 mm (fig. 3a). The bands consist of alternating layers of a highly porous spongy tufa from 0.3 to 2 mm thick and tan-colored bands of a more dense tufa from 75 to 200 microns thick. Thin crenulated laminae (20 to 25 μ) of dark-brown dense tufa mark the transition between the lobate tufa and the underlying tubular structure. Pockets in the spongy tufa forming the lobes are filled or partly filled with calcareous oölites and clastic mineral debris cemented by microcrystalline calcite. Plagioclase, green hornblende, and biotite are the most common noncalcareous minerals. The oölites, which are 0.25 to 1.0 mm in diameter, have nuclei of plagioclase or opaque minerals.

## Pinnacle Structure

*Tower and tombstone pinnacles.*—Most of the tower and all of the tombstone-shaped pinnacles of the northern group consist of an inner core of horizontally banded stony lithoid tufa and an outer layer of banded cavernous lithoid tufa (figs. 7 and 8). The outer cavernous tufa tends to break away from the pinnacles and consequently the inner stony lithoid cores of most of the pinnacles are exposed. On the few pinnacles which still retain much of their outer cover, the cavernous tufa is as much as 10 ft thick along the sides of the pinnacles and 10 to 20 ft thick at the summits of the pinnacles (fig. 8).

Horizontal and vertical fractures or joints have developed in the stony lithoid cores and consequently large columnar sections have toppled from the summits of many of the pinnacles, causing them to become flat-topped.

The bases of most of the northern tower- and tombstone-shaped pinnacles are encircled by 2 to 5 ft of vertical layers of nodose and tubular tufa which pass beneath the gently sloping lacustrine sediments flanking the bases of the pinnacles (fig. 7). A single adequate exposure of the flanking basal slopes shows that immediately adjacent to

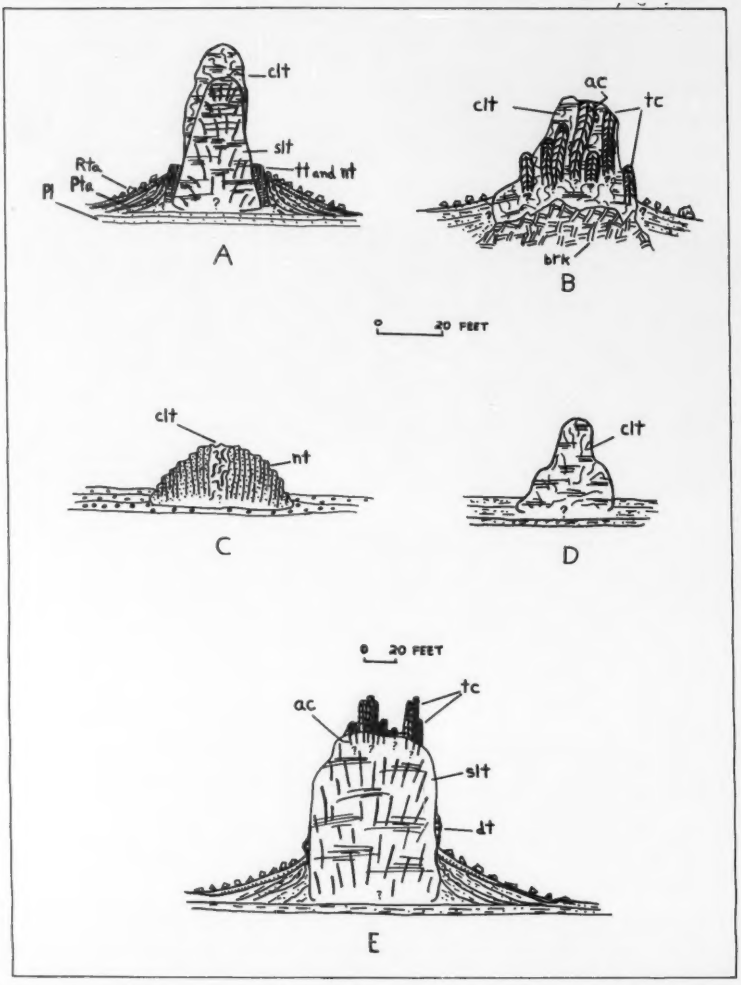


FIG. 7.—Schematic vertical sections showing structures of pinnacles at the southwest end of Searles basin. A—tower-shaped pinnacle of northern group; B—tower-shaped pinnacle along southern margin of northern group and of middle group; C—cone-shaped pinnacle along northern margin of northern group; D—tower-shaped pinnacle of southern group; and E—end view of northern limestone ridge. Symbols read as follows: clt—cavernous lithoid tufa, slt—stony lithoid tufa, nt—nodose tufa, dt—dendritic tufa, tt—tubular tufa, tc—tubular columns, ac—axial channels in tubular columns, brk—igneous and metamorphic bedrock, Pl—Pleistocene lake sediments, Pta—Pleistocene sublacustrine talus, and Rta—Recent subaerial talus.

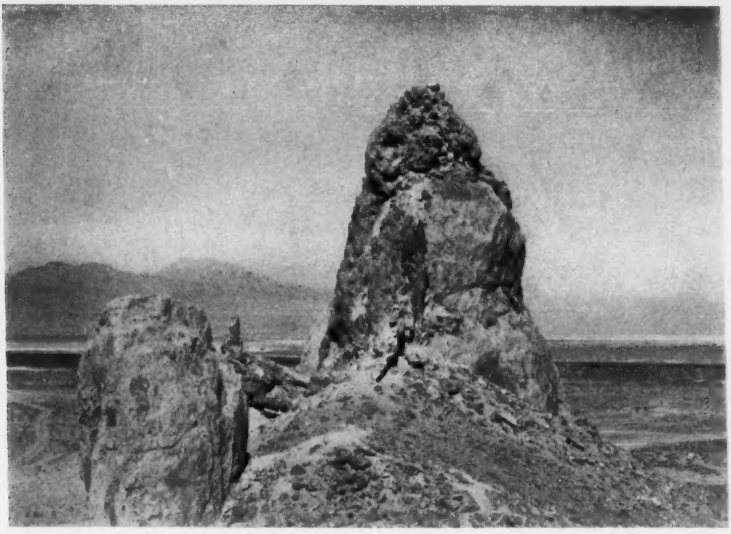

FIG. 8.—Stony lithoid inner core of a tower-shaped pinnacle of the northern group (light color), which still retains a portion of the outer cavernous tufa layer (dark color). Scale is shown by figure standing (arrow) at base of pinnacle; the salt-encrusted surface of Searles Lake can be seen in background.

the pinnacles the lacustrine sediments consist of steep outward-dipping crenulated layers of sandy calcareous silt. Radially outward the sediments dip less steeply and gradually merge with the nearly horizontally bedded marly silts and sands which surround and underlie the pinnacles (fig. 7). A layer of dense sandy tufa, which is overlain by about 1 ft of loose talus derived from the pinnacles, forms the upper 1 to 2 ft of the flanking slopes.

Near the southern margin of the northern pinnacle area and throughout the middle pinnacle group, tower-shaped pinnacles lack inner stony cores and are constructed of cavernous lithoid tufa surrounding and overlying numerous vertical columnar bodies of tubular tufa (figs. 7 and 9). Tubular columns range from about 4 in to 6 ft in diameter and are near the central part of the pinnacle. The outer few inches of the columns consist of banded lobate tufa which gives the exterior surface of the columns a distinct mammillary appearance. The bases of the columns were nowhere observed, but the exposed segments are usually not more than 5 ft high, and only a few of the columns reach the summits of the pinnacles (fig. 9). Each column has a 3 to 6 in axial channel of exceptionally porous and honeycombed tufa; these channels terminate from 6 to 8 in below the tops of the columns (fig. 7).

Tower-shaped pinnacles of the southern group are mostly horizontally banded masses of cavernous lithoid tufa and some dendritic tufa. The pinnacles are greatly weathered and the tufa is highly fluted and reddish-orange in color. Many of these pinnacles have unfortunately been destroyed by blasting.

*Cone-shaped pinnacles.*—Cone-shaped pinnacles of the middle and southern groups are mostly structureless mounds of cavernous lithoid tufa. However, cone-shaped pinnacles forming the lakeward or northern periphery of the northern group consist of a central core of cavernous lithoid tufa surrounded by vertical layers of nodose tufa (fig. 7). The central cores range from 2 to 8 ft in diameter and include many narrow

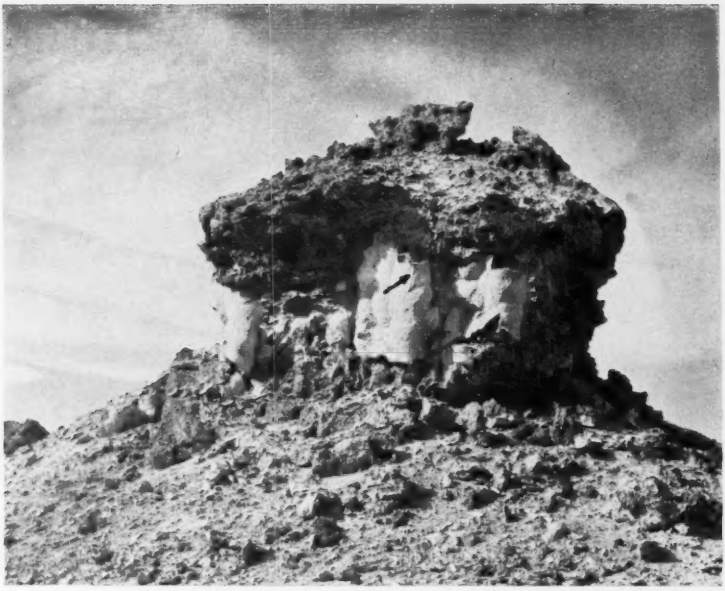

FIG. 9.—Exposure of chalk-white weathering columns of tubular tufa in a tower-shaped pinnacle of the southern group. Hemispherical lobate tufa forms the outer layer of the columns; cavernous lithoid tufa surrounds and overlies them. Note prospector's pick for scale (arrow).

vertical passages. Circumscribing layers of nodose tufa are 4 to 8 in thick, but their aggregate thickness is 20 to 35 ft.

*Limestone ridges.*—Stony lithoid tufa forms the bulk of the northern limestone ridge. The ridge is castellated with summit pillars of tubular tufa which rise from 5 to 15 ft above its average height (figs. 2 and 7). Each pillar comprises as many as 10 to 15 columns of tubular tufa. The diameters of the columns range from a few inches to about 4 ft. Hemispherical lobate tufa does not form the outer layer of the columns atop the northern ridge but is about midway between the axial channel and the outer surface of the column. Axial channels are exposed at the tops of many of the larger columns.

The two limestone ridges in the middle group lack summit pillars and apparently consist of outward-dipping hummocky layers of massive lithoid tufa. The summits and sides of the ridges are coated with swirling masses of a more porous tufa. Many cracks, joints, and voids in these ridges are filled or lined with fresh-appearing tubular and digitate tufa. The largest of the two ridges is 800 ft long, 500 ft wide, and 140 ft high; it is a very conspicuous feature and can be seen from a distance of several miles (fig. 10). A number of smaller outlying mound-like pinnacles of massive lithoid tufa are closely associated with both of the ridges.

Although the exact field relationship of the two ridges to the other pinnacles of the middle group is not fully known, it is tentatively concluded that the two ridges and the associated smaller mounds are older and formed under different conditions than the other pinnacles. This conclusion is based upon (1) the exceptional denseness of the massive lithoid tufa forming the ridges and the outlying mounds, (2) their lack of structures common to the other pinnacles, (3) the presence of unweathered tubular and digitate tufa in joints and cavities, and (4) the fact that the ridges are by far the largest of the pinnacles in Searles Basin.

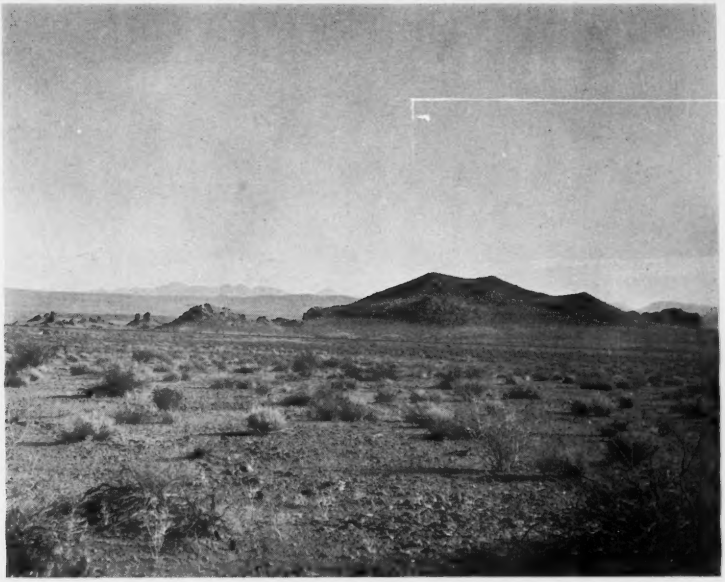

FIG. 10.—On the right is the largest (140 ft high) of the two limestone ridges of the middle group. Smaller cone- and tower-shaped pinnacles are in the left background; view is to the southeast.

*Buried tufa deposits and pinnacle foundations.*—Along the northern (lakeward) periphery of the northern pinnacle area, the upper 40 to 50 ft of the fine-grained to gravelly lacustrine sediments which underlie the pinnacles is exposed in stream cuts. Within this section are buried tufa deposits which must be somewhat older than the pinnacles (fig. 11). Stream erosion has also exposed a number of buried deposits farther to the south within the major area of the northern pinnacles. An excellent exposure is about half a mile northwest of the pinnacles where a railroad cut exposes a large structureless mound of tufa buried beneath an ancient gravel bar.

The buried tufas are mostly bedded lens-shaped deposits from 5 to 15 ft thick (fig. 11). They accumulated on gravelly sediments and consist of stony lithoid tufa; their weathered surface is highly fluted and commonly breaks in jagged icicle-like pieces.

Although the buried tufas must be older than the overlying pinnacles, field relations indicate that the two deposits are penecontemporaneous. The buried bodies closely underlie the tower-, tombstone-, and ridge-shaped pinnacles and conceivably some of the older deposits served as foundations for the construction of these pinnacles. However, the small cone-shaped masses of nodose tufa along the northern margin of the pinnacle area do not have deep subsurface extensions and are deposited on beach gravels directly overlying the older buried masses. Tower-shaped pinnacles of cavernous and tubular tufa near the southern margin of the northern pinnacle area also do not have deep subsurface extensions and have formed on outcrops of basement rock (fig. 7).

Buried tufa lenses were not observed in the lacustrine sediments underlying the pinnacles of the southern and middle groups, and presumably these pinnacles do not have extensive subsurface segments. The bases of a few of these pinnacles are exposed, and typically their foundations consist of horizontally bedded cavernous lithoid tufa. Some of the pinnacles along the northwest side of the southern group have formed on

FIG. 11.—Exposed bedded lenticular mass of tufa in the lacustrine sediments which underlie the northern pinnacles. In the background are a number of small cone-shaped pinnacles which form the northern or basinward periphery of the northern pinnacle group.

outcrops of basement rock, but most of these pinnacles rest upon lacustrine sediments.

## ORIGIN OF SEARLES LAKE PINNACLES

### Knowledge of Tufa Deposition

The origin of tufa is by no means a point of common agreement among geologists. Suggested origins have varied from petrified mushrooms to the partial product of bacteria, but only two general proposals have had much support. These are: (1) tufa is primarily an inorganic chemical precipitate; or (2) tufa is primarily an organic deposit precipitated by algae.

After the Pleistocene tufas of Lake Lahonton, Mono Lake, and Lake Bonneville were studied by King (1878), Russell (1885 and 1887), and Gilbert (1890), the general concept prevailed that tufa was a chemical precipitate from saturated lake water. Dunn (1953) has more recently reproposed a chemical origin for the Pleistocene and presently forming tufas at Mono Lake, California.

One of the first to suggest an algal origin for deposits similar to tufa was Weed (1889), who investigated the travertine spring deposits at Yellowstone National Park. Upon reviewing Weed's work, Russell altered his opinion and favored an algal origin for the tufa deposits at Lahonton and Mono Lakes (Russell, 1893, p. 94–95).

Subsequently, Jones (1914 and 1925) offered evidence that the modern and Pleistocene tufa deposits of the Salton Sea and Lake Lahonton were precipitated by algae. His principal evidence was:

(1) the presence of algal remains in the Pleistocene tufas;

(2) the tendency of modern and Pleistocene tufas to develop toward the direction of sunlight;

(3) the abundance of snails and other animal life associated with the Pleistocene tufa suggests a food supply furnished by algae;

(4) recent tufas can be shown to be a precipitant of algae.

Bradley (1928) later made a comprehensive investigation of the fresh-water limestone reefs of the Eocene Green River Formation. He was able to demonstrate that the reefs were largely precipitated by algae, in particular by the species *Cylorellopsis coloniata* Reis. Bradley (1928) also described presently forming algal reefs at Green Lake, New York.

After Bradley, Howe (1931) discussed the general geologic importance of lime-secreting algae and described a new travertine-forming organism. More recently Johnson (1943, 1954, and 1957) reemphasized the significance of algae as rock-forming organisms and described (1937) a fresh-water algal tufa of Oligocene age from Colorado.

## Deposition of Searles Lake Tufa

In thin section, stony and cavernous lithoid tufa and dendritic tufa are seen to contain abundant small ellipsoidal to polygonal bodies which are believed to be molds of algal cells (figs. 4, 5, and 6). The diameters of the cells range from about 60 to 130 $\mu$ and average approximately 80 $\mu$. Cells and associated algal structures constitute the bulk of the lithoid tufas, but dendritic tufa consists primarily of alternating light and dark layers of non-cellular tufa (fig. 5).

Optically oriented calcite, possibly aragonite, fills many of the cells, but many are rich in organic matter and are dark brown. Most of the cells are encircled by dark-brown bands containing organic matter which closely follow the configuration of the cell (figs. 4 and 6). Where cells are closely grouped, the outer banded zone surrounds the entire cluster rather than the individual cell. Owing to the organic matter, which by ignition loss at 450° C averages 1.1 and 2.4 percent for stony and cavernous lithoid tufa and 3.2 percent for nodose tufa, the algal cells and the surrounding dark bands can be etched into relief with weak acetic acid. Careful etching of impregnated thick sections will leave botryoidal clusters of cells attached to residual masses of the impregnating medium.

Both the mega- and microsopic structures of the Searles tufas are similar to fresh-water algal limestones described and figured by Jones (1914 and 1925), Bradley (1928), Howe (1931), Johnson (1937, 1943, and 1957), and Hill and Tompkin (1953).

An additional character of the Searles tufas which may indicate an organic origin is the upward and outward radiating structure of the tubular columns. This structure is easily explained if the tubules were constructed by algae attempting to present maximum surface areas to available nutrients and solar radiation.

Algal cells were not observed in nodose and lobate tufas, but their microbanded concentric structure suggests that they may be partially of algal origin. Neither concentric bands nor cells were noted in the massive lithoid tufas of the ridges of the middle group. It may be that these ridges and the associated outlying mounds were largely built by inorganic processes.

Undoubtedly some of the tufa in the Searles pinnacles is of inorganic origin. This origin is indicated by the bands of relatively coarse-crystalline calcite which fill or line many of the cavities in the porous framework of the tufa. However, the fact that stony and cavernous lithoid tufas, which constitute between 50 and 95 percent of most of the pinnacles, consist principally of algal structures indicates that algae were the principal agents causing deposition of the tufa. It is, of course, difficult to maintain that algae were essential to the precipitation of the tufa. Very probably calcium carbonate would have accumulated without the aid of algae, though possibly not in the form of pinnacles.

It is not fully understood by the writer, nor apparently by others who have studied Pleistocene pluvial tufas, why the several varieties have formed. This is particularly true for those Pleistocene lake basins which have deposits of thinolite tufa, a little understood variety (Radbruch, 1957). One can only surmise that the chemistry of the water, the growth habits of algae, and the relative amount and rate of contemporaneous inorganic and organic precipitation were involved in fashioning in variations at Searles Lake.

## Deposition of the Pinnacles

The pinnacles are thought to have formed subaqueously. This interpretation is based on the facts that (1) ostracod valves are found in the tubular tufa on top of the northern limestone ridges, which is the tallest of the northern pinnacles; (2) none of

the pinnacles rise above the highest shoreline left by the Pleistocene lakes; (3) the pinnacles are underlain and flanked by lacustrine sediments; and (4) much of the tufa is believed to have been precipitated by aquatic plants.

Axial channels in the tubular columns, the porous and cavernous nature of the tufa surrounding and overlying the columns, and the pinnacled form of the tufa deposits are thought to indicate that the pinnacles were precipitated about the orifices of ascending sublacustrine springs.

The lineations of the pinnacles probably reflect the strikes of faults in the igneous and metamorphic rocks which underlie the lacustrine sediments. These presumed faults are probably structurally related to the Garlock Fault, one of the major structural elements of southern California, which passes in a northeasterly direction along the southern margin of Searles Basin within a few miles of most of the pinnacles.

It is both of interest and of significance that Russell (1885 and 1887) found springs rising through the tubular and porous centers of submerged pinnacles at Mono Lake. Moreover, other pinnacles along abandoned shorelines at Mono Lake are known to be aligned along faults and to be the loci of springs (Russell, 1885, 1887; Dunn, 1953).

The various shapes of the Searles pinnacles probably reflect the size, flow, and chemical composition of the springs, the spacing of the springs, the depth of water, and the rate at which the lake surface was rising or falling. Tombstone and ridge structures may have accrued about a number of closely spaced springs of considerable flow. Tower- and cone-shaped pinnacles probably developed about one or only a few springs. Many of the cone-shaped pinnacles of the northern group rest upon beach gravels, and it is apparent that they were precipitated in shallow water and were among the last of the pinnacles to form. Virtually all of the buried lenticular bodies grew upward from gravel lenses, and these tufas probably also formed in nearshore areas, although it is not clear that springs were directly involved in their precipitation.

## Age of the Pinnacles

Inasmuch as the pinnacles are thought to have accumulated subaqueously, the time of accumulation was doubtless during the Pleistocene when the basin was periodically filled with rain, runoff, and glacial melt water (Gale, 1915; Blackwelder, 1954; Smith and Pratt, 1957).

During this epoch at least four separate periods of glaciation were recorded in the Sierra Nevada (Blackwelder, 1931). Any of these glaciations could have brought about a flooding of the basin with subsequent precipitation of the pinnacles. However, the pinnacles of the northern and middle groups are in general structurally similar and rest upon or slightly within the youngest lacustrine sediments at the southwest end of the basin. These pinnacles are also fresh and unweathered in appeareance. It is therefore thought that they formed in the last major fresh-water lake occupying the basin. Conversely, the pinnacles of the southern group contain none of the structures of the middle and northern groups and are deeply weathered and decrepit in appearance, a fact previously emphasized by Blackwelder (1941). For these reasons the southern pinnacles are thought to be older than the other groups and to have accrued in a preceding and necessarily somewhat deeper lake as the southern pinnacles are at higher elevations than the middle and northern groups.

Radiocarbon dates of two layers of organic mud, the Parting Mud and the Bottom Mud, which are beneath the salt-encrusted surface of Searles Lake and separated stratigraphically by 54 ft of evaporite salts (Haines, 1959), indicate that the last major fresh-water lake occupied the basin between 25,000 and 10,000 years ago and that a preceding fresh-water lake filled the basin prior to 32,000 years ago (Flint and Gale, 1958; Smith, 1958). Smith correlated the time of deposition of the upper mud layer, the Parting Mud, with Blackwelder's late Pleistocene Tioga glacial stage, and the Bottom Mud with his Tahoe stage. Blackwelder (1941) earlier recognized the two-stage late Pleistocene lake history of Searles Basin and assigned a Tioga age to a younger and smaller lake and a Tahoe age to a preceding deeper lake which overflowed to the southeast into adjacent Panamint Valley. The results of the present study, which corroborate earlier opinions of Blackwelder (1941 and 1954), indicate that the pinnacles of the northern and middle group formed in a

geologically recent lake probably of Tioga age between 25,000 and 10,000 years ago and that the southern pinnacles precipitated in an older and deeper lake most likely of Tahoe age which filled the basin probably sometime between 100,000 and 32,000 years ago.

The two limestone ridges of the middle group and the associated smaller outlying mounds of massive lithoid tufa may have accrued in the Tahoe lake or possibly they were precipitated as subaerial deposits during the inter-pluvial period separating Tahoe and Tioga Searles Lake.

## DEPOSITIONAL HISTORY AND INFERRED HYDROLOGY

Radiocarbon dates (provided that they prove applicable) would establish the depositional history of the pinnacles and would afford a tool for reconstructing the sequence of tufa deposition and certain phases of Tahoe and Tioga hydrology at the southwest end of the basin. By acknowledging this fact it may seem premature to speculate on these events when only physical evidence is used as a guide. Nevertheless, the following sketch of the depositional history of the pinnacles and the implied hydrology is offered as a start toward a more comprehensive understanding of the tufa deposits in Searles Basin.

### Maximum Tahoe and Tioga Lake Depths

Tahoe Searles Lake is known through several lines of evidence to have overflowed at an elevation of 2260 ft into adjacent Panamint Valley (Gale, 1915; Blackwelder, 1941 and 1954, and Hubbs and Miller, 1948). Pinnacles of the southern group which are believed to be of Tahoe age reach but do not surpass this elevation. The base of the Bottom Mud, which is of Tahoe age and approximately 100 ft thick, is about 860 ft lower than the 2260-ft overflow point. Hence if post-Tahoe compaction of lake sediments and basin subsidence have not been excessive, and if Tahoe Searles Lake overflowed early in its history, then the lake attained a maximum depth near 860 ft.

The maximum depth of the Tioga lake is more difficult to reconstruct because of the lack of a clearly recorded upper shoreline for this lake. However, the pinnacles of the middle group, which are considered to be of Tioga age, attain an elevation near 2000 ft. This elevation, therefore, establishes a threshold level reached and probably somewhat surpassed by the Tioga lake if it is reasoned that the pinnacles would have tended to remain near the lake surface in the critical euphotic zone. The difference in elevation between the threshold level and the base of the Parting Mud, which is of Tioga age and about 12 ft thick, is about 460 ft. A depth of 460 ft is, in the author's opinion, probably close to the maximum depth attained by Tioga Searles Lake.

### Sequence of Pinnacle and Tufa Deposition

Pinnacles of the southern group are essentially structureless masses of lithoid tufa, and very little can be stated regarding their depositional sequence. It seems clear, however, that they formed about springs issuing along faults striking N. 65° W., N. 50° W., N. 55° E., and N. 30° E. when the lake was at or near its overflow point.

Pinnacles of the northern and middle group are composed of several varieties of tufa, and it is possible to speculate more freely on their depositional sequence. This is attempted below.

It is evident from radiocarbon dates of the Parting Mud that Tioga Searles Lake began to form about 23,000 years ago (Flint and Gale, 1958; Smith, 1958). However, it was not until the lake had attained a depth of about 200 ft that tufa deposition began at the southwest end of the basin (table 2). These initial deposits were formed by algae on nearshore gravels as bedded lens-shaped deposits of stony lithoid tufa. Most of these early accumulations were subsequently buried by clastic sediments as the lake level continued to rise; however, some may have continued to grow upward to eventually form pinnacles.

At nearly the same time, algae precipitated additional stony lithoid tufa about the orifices of sublacustrine springs issuing primarily along faults striking N. 30° E. Precipitation of lithoid tufas about the springs was not suppressed by contemporaneous clastic sedimentation, and for some time the tufas continued to build upward toward the rising lake surface. After these pinnacles attained their maximum diameters, nodose and tubular tufas were deposited about their bases and sandy to silty

TABLE 2.—*Sequence of pinnacle and tufa deposition and inferred Tioga hydrology*

| Pinnacles and tufa deposited | Approximate depth of lake during deposition (1540 ft taken as average elevation of Parting Mud). | Approximate elevation of lake surface. | Age |
|---|---|---|---|
| Buried lenses of stony lithoid tufa | 170 to 210 ft (rising) | 1710 to 1750 ft | Somewhat younger than 23,000 years old |
| Northern pinnacles; stony lithoid tufa, some nodose and tubular tufa | 210 to 410 ft (rising) | 1750 to 1950 ft | ? |
| Middle pinnacles; cavernous lithoid, tubular and lobate tufas | 380 to 460 ft (rising to maximum, then falling) | 1920 to 2000 ft | ? |
| Northern pinnacles; cavernous, Tubular, and lobate tufas | 410 to 190 ft (falling) | 1950 to 1730 ft | ? |
| Northernmost cone-shaped pinnacles; cavernous lithoid and nodose tufas | 220 to 180 ft (falling) | 1760 to 1720 ft | Probably about 11,000 to 12,000 years old |

calcareous sublacustrine talus sediments farther out.

Upon reaching a depth of approximately 300 ft, the lake waters flooded an arm-like bay at the southwest end of the basin. Continued deepening of the lake extended this bay farther to the southwest and the cavernous lithoid, dendritic, and tubular tufas of the pinnacles of the middle group formed about springs issuing along faults primarily striking N. 30° E. At its maximum stand, when the lake was somewhat deeper than 460 ft and stood at an elevation near 2000 ft, the youngest pinnacles at the extreme southwest end of the bay were probably within a few feet of the surface, but most of the pinnacles of the northern group were now submerged beneath 160 ft of water; however, a few were within 60 to 100 ft of the surface. Apparently the lake level rose too rapidly during the flooding of the bay for many of the northern pinnacles to maintain themselves within the critical euphotic zone. The two older limestone ridges of the middle group were only partly submerged when the lake stood at its maximum stand and probably at this time tubular and digitate tufas were deposited in joints and voids in their structure. The swirling masses of tufa coating the summit and sides of the ridges may have formed in spring water rising and escaping through the ridges.

During the subsequent falling of the lake level, deposition of cavernous, dendritic, and tubular tufas continued, and possibly additional cone- and tower-shaped pinnacles accumulated as the lake withdrew from the bay. When the lake level had dropped to within a few tens of feet of the summits of the northern pinnacles, cavernous and tubular tufas were pecipitated over the pinnacles and subsequently the sandy dense tufa which overlies the flanking fine-grained lacustrine talus sediments. During the final withdrawal of the lake from the northern pinnacle area, which possibly took place between 12,000 and 11,000 years ago, small cone-shaped pinnacles of nodose and lithoid tufa formed upon beach gravels overlying the older and now completely buried initial tufa deposits.

The absolute dating of the Searles pinnacles by the radiocarbon method would, of course, add immeasurably to the generalized sketch given above, since it would provide a means for more accurately interpreting the depositional history. Moreover, radiocarbon dating of other tufa deposits at Searles Lake should allow time correlations with the subsurface sediments and thus provide a better understanding of the late Pleistocene hydrology at Searles Basin.

ACKNOWLEDGMENTS

The manuscript was read by Dr. George I. Smith of the U. S. Geological Survey and by Drs. Kenneth O. Emery, Richard O. Stone, and Mr. Ian Kaplan of the University of Southern California. The author is grateful for their many helpful suggestions. Dr. Smith gave freely of much of his time to discuss various phases of the study. Phillip L. Scholl, Alden R. Tagg, and Kathlene M. Tagg assisted the writer in much of the field work. Drs. Thomas Clements and William H. Easton of the University of Southern California originally encouraged the investigation.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 432–446
FIGS. 1–5, SEPTEMBER, 1960

# ORIGINAL BEDROCK COMPOSITION OF WISCONSIN TILL IN CENTRAL INDIANA[1]

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## ABSTRACT

An attempt has been made to convert the mineral and rock-fragment components of a till into the original bedrock materials from which they were derived. The till used was a composite one (designated the "typical" till) that represented the averaged fraction-analysis data for 11 highly similar till samples from Marion County, central Indiana. For this study it was assumed that: (1) the glacier flow unit that reached Marion County was 20 miles wide and coincided with a series of interconnected bedrock lowlands lying between northeastern Indiana and Lake Naococane, Quebec, (2) the flow unit eroded bedrock (or older drift) continuously throughout its journey from Lake Naococane, and (3) the different shales traversed by the ice were eroded equally.

In terms of volume-percent composition, the typical Wisconsin till in Marion County probably consists of about 33 percent shale, 29 percent metamorphic and igneous rock, 18 percent limestone, 16 percent dolomite, 2 percent sandstone and siltstone, and 2 percent miscellaneous rocks. Approximately 90 percent of the typical till may consist of bedrock from outcrops more than 100 miles upstream from the site of till deposition.

An approximate 1 to 1 relationship has been found between the weight percent of a given rock type in the typical till and the percent of outcrop area of that rock type upstream. Implications of this relationship are that: (1) the ice streams that developed the Marion County tills eroded materials according to a weight rule, (2) the tills were not deposited by lodgement processes, and (3) the ice streams picked up and transported bedrock for distances in excess of 1200 miles.

## INTRODUCTION

### The Typical Marion County Till

Fraction analyses (mineralogy and lithology against size) were made recently (Harrison, 1959) for 11 samples of till of Wisconsin (and Illinoian?) age from Marion County, central Indiana (fig. 1). The samples were taken from three or four different till sheets, and the sampling localities (fig. 1) fell within an area of 250 square miles. Detailed statistical analyses (Harrison, 1959 p. 17), of the laboratory data indicated that the samples were so greatly similar in terms of both texture and composition that they might well have been subsamples of a single till sample. For this reason, it was considered acceptable to average the data from the 11 analyses for concise presentation. The averaged data are presented in figure 2, a "texturo-compositional" diagram. Although two of the till samples may be Illinoian in age (Harrison, in preparation), figure 2 depicts what may be termed the "typical" till of *Wisconsin* age from Marion County, Indiana (fig. 1). As might be inferred from figure 2, the percentages of cobbles and boulders in the typical till are sufficiently small (Harrison, 1959, table 4) that they may be ignored.

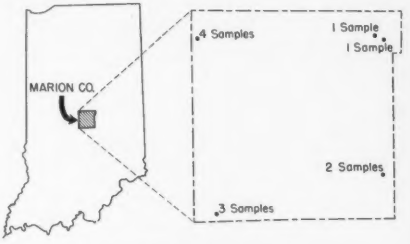


FIG. 1.—Marion County, Indiana, and sampling localities.

### The Problem

This study was initiated in an attempt to answer the question: *What is the original bedrock composition of the typical Marion County till of Wisconsin age?* How much shale bedrock, for instance, is represented

[1] Published with the permission of the Acting State Geologist, Indiana Department of Conservation, Geological Survey. Manuscript received November 9, 1959.

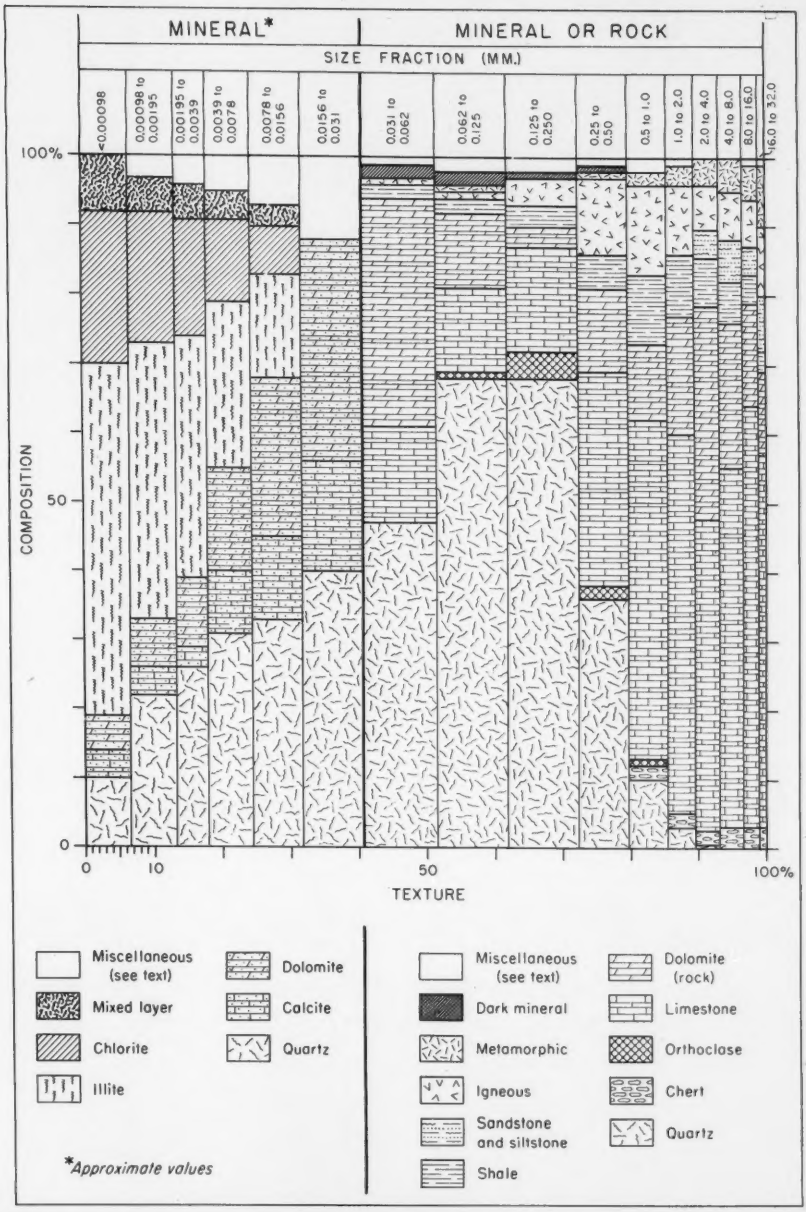


FIG. 2.—Texturo-compositional diagram depicting typical Marion County till of Wisconsin age. (Compositional values for fractions <0.031 mm good to ±30 percent for constituent shown.)

by the clay and associated minerals in the fine fractions (fig. 2) of the till? How much orthogneiss is represented by the grains of quartz, feldspar, and accessory minerals in the sand and silt fractions? If a process could be developed for converting mineral components into bedrock types, one could add the values obtained by the conversion process to those of the rock materials known to be present and arrive at the total percentages of the various rock types in the till. A knowledge of the percentages of the original bedrock types in a till would permit one to make quantitative speculations about the genesis of the till.

Many assumptions must surround an attempt to answer the question of the foregoing paragraph. The most critical assumptions are related to the characterization of the ice stream that developed the Marion County till. Throughout the study the author will attempt to point out alternatives in the methodology employed. He wishes to stress the tentative nature of the final quantitative results.

## APPROACH

Percentages of original bedrock materials in a till can be estimated from a knowledge of two related distributions: (1) the distribution of the rock types that underlie the bedrock surface from which the till was derived and (2) the distribution of the mineral species in each of the various bedrock types of this surface. A knowledge of the distribution of rock types at the bedrock surface (point 1) involves at least two assumptions. The first assumption is the flow-path taken by the ice stream that developed the till. A second, and equally important assumption, involves the upstream and downstream limits of bedrock erosion along the assumed flow path. The distribution of mineral species in each of the bedrock types along the assumed flow path (point 2) can be approximated reasonably well by laboratory studies of samples of the rocks or by reference to published analyses of the rock types.

Points of departure in this study include (1) the basic data of till composition (fig. 2) and (2) a study by Horberg and Anderson (1956) on the origin of Pleistocene glacial lobes and ice streams and their movement into the central United States. The following paragraphs elucidate each of the steps in the study.

## PROPOSED FLOW PATH

Horberg and Anderson (1956) recognized lines of flow in the Wisconsin ice sheet that caused the major lobes and sublobes that moved into the Central Lowland. They postulated (1956, table 2; p. 112) that central Indiana was glaciated during the Tazewell Substage by ice from the "Ontario-Erie glacier" and that this ice originated in the Labradorean center of glaciation.

On the basis of Horberg and Anderson's study and the *principle* that an advancing ice sheet tends to flow parallel to broad, elongate bedrock lowlands (Horberg and Anderson, 1956, p. 109), it is possible to propose a flow path for the ice that moved out of the highlands of central Quebec southwestward to central Indiana. This proposed flow path is bounded by the dashed lines shown on figure 3. The bounding lines coincide with the 750-foot bedrock contour in the vicinity of the lake basins but are of arbitrary width elsewhere. The flow path outlines a series of interconnected bedrock lowlands, and localization of relatively rapidly moving ice along the center line of this path resulted in what may be termed the Ontario-Erie *flow unit* (see p. 437) of the Wisconsin ice sheet. The course taken by the part of the Ontario-Erie flow unit that reached Marion County is drawn (figure 3) slightly more than 20 miles in width; this is a highly idealized picture of the ice stream that deposited the till in this central Indiana county.

A recent study by Dreimanis and Terasmae (1958) provides substantiating evidence for portions of the flow path shown on figure 3. Dreimanis and Terasmae (1958, fig. 2, B) indicate that the ice that deposited the basal (classical) Wisconsin tills in the Central Lowland may have taken a path in the Lake Ontario region identical to the one shown for the flow unit on figure 3. They further indicate, as does Anderson (1957, p. 1427), that the ice of the Ontario-Erie lobe that also deposited the early Wisconsin tills northeast of Marion County (Anderson, 1957, p. 1421) traversed the Grenville Series of southern Quebec and southwestern On-

tario. It is precisely such a course that is shown on figure 3.

### "VERIFICATION" OF FLOW PATH

No known indicator stones were found in the samples of till from Marion County; not a single fragment of jasper conglomerate (Slawson, 1933), for instance, was encountered. (Lack of jasper conglomerate constitutes negative evidence for ice flow out of the Lake Huron basin.) From an examination of the flow path (fig. 3), one could say that the metamorphic rock fragments encountered in the till should be siliceous and almost exclusively rich in potash feldspar (cf. Adams, 1896, p. 38; fig. 3 this study). This is certainly true for the great majority of metamorphic fragments studied, and it is significant that in counts of more than 11,000 sand and very coarse silt grains, only 12 plagioclase grains were identified.

Ratios of purple to red garnets in the heavy-mineral suites of the tills provide information as to the flow path taken by the ice (Dreimanis and others, 1957). Ratios of purple (and faintly pink) to reddish-brown garnets (1.9 to 1) in the typical till indicate glacier flow through the region north of Ottawa but not farther west (Dreimanis and Terasmae, 1958, p. 126). This ratio is sensitive to the operator's concept of "red" and "pink"; therefore, it was determined for the author by A. Dreimanis (1959, written communication) from a slide of the heavy minerals from the typical till of Marion County.

Among the prominent sedimentary rock fragments in the typical till (fig. 2) that have an origin in Canada are graywackes[2] and "tillites" (also termed "pebble conglomerates" or "pebbly graywackes" by different authors). These types may be present along the proposed flow path (Wilson, 1913, p. 129; Mawdsley and Norman, 1935, p. 43). Seven quartz pebbles, 8–16 mm in size, were found among approximately 3500 pebbles identified during fraction analysis. They fit the description of pebbles found in rock types from the locality of the flow path in south-central Quebec (Bergeron, 1957, p. 129). Pebbles of quartz might also have been derived from outcrops of the Sharon Conglomerate in northeastern Ohio, although these outcrops are presently on high ground beyond the limits of the flow path (fig. 3).

Table 1 is a breakdown of the percentages of various bedrock types of Canadian derivation that are found in the granule and pebble sizes of the typical till. (A few of the sandstone fragments may have been derived from outcrops in Indiana.) Although the breakdown of table 1 indicates little in a specific way about the flow path in question, the following observations indicate that the positioning in Canada (fig. 3) is reasonable:

1. *Presence of a large percentage of phaneritic acid-igneous fragments.* It is believed that most of these "igneous" fragments are amenable to reclassification as "orthogneiss" fragments. Adams (1896, p. 38), for example, devotes a section of his report that deals with rock types marginal to the flow path north of Ottawa to "Gneisses of Igneous Origin." He stated (p. 31) that foliation in the orthoclase gneiss of the area "may be so indistinct that its existence can only be detected by the examination of large weathered surfaces." The "granitic" rock fragments of table 1 contain potash feldspar almost exclusively. A distinct lack of plagioclase grains in the sand fractions of the till further indicates that the ice traversed the orthoclase-gneiss areas that were described by Adams as extending into the area of the flow path north of Ottawa (fig. 3).

2. *Presence of a few porphyritic fragments.* These pebbles and granules could be derived from a bedrock area like that in the vicinity of Lake Mistassini (fig. 3) where "intermediate and basic flows and sills and their metamorphic equivalents" are present (Quebec Department of Mines, 1957, Geological Map of Quebec). There are other sources outside of the flow path, however.

3. *Presence of quartzite.* Metaquartzite fragments that are found in the Marion County till could have been derived from the area between about the 46th and 51st parallels (Bergeron, 1957, p. 127; fig. 3) and from outcrops south-southwest of Ottawa (Wilson, 1946, map 852A). Positioning of the flow path (fig. 3) much farther

[2] Anderson (1957, table 8) did not recognize "tillite" or graywacke pebbles in the drift deposited by the Ontario-Erie lobe in northeastern Indiana. Both of these types, however, are present (W. J. Wayne, personal samples) in Anderson's so-called "index" grade size (1957, p. 1417).

TABLE 1.—*Rock fragments 2–32 mm in size largely or exclusively of Canadian derivation*

| Rock Type[1] | | Percentage of All Types Listed |
|---|---|---|
| *Sedimentary:* | | |
| Sandstone (exclusive of types below) | | 12 |
| Arkose | | 0.5 |
| Graywacke | | 12 |
| "Tillite" | | 3 |
| *Igneous:* | | |
| Acidic: (Granite to monzonite) | Aphanitic | 7 |
| | Porphyritic | 2.5 |
| | Phaneritic | 28 |
| Basic: (Granodiorite to gabbro) | Aphanitic | 2 |
| | Porphyritic | 0.5 |
| | Phaneritic | 4 |
| Ultrabasic: | | 0 |
| *Metamorphic:* | | |
| Light | Massive or Granulose (mostly quartzite) | 8 |
| | Lineate or Foliate | 8 |
| Inter. (incl. red or brown) | Massive or Granulose | 2 |
| | Lineate or Foliate | 4 |
| Dark (incl. green) | Massive or Granulose | 2 |
| | Lineate or Foliate | 5 |
| Total | | 100.00 |

[1] Terminology after Travis (1955).

to the west would limit the contact of the ice with quartzite bedrock.

4. *Presence of arkose pebbles.* Quartzose and arkosic sandstone pebbles that were found in the typical till are similar in appearance to types described (Bergeron, 1957, p. 127) from bedrock areas underlying the flow unit in Quebec (fig. 3). Four pebbles of arkose were found in counts of about 3500 pebbles of all types. Mawdsley and Norman (1935, p. 43) report arkose at Chibougamau Lake.

Figure 3 shows a few patches of "layered anorthosite complex" in the vicinity of the 50th parallel. It is perhaps significant that about 10 anorthosite pebbles were found in counts of approximately 3500 pebbles of all types from the 11 till samples. Once again, however, there are other sources of anorthosite outside of the flow path (fig. 3).

Other prominent sedimentary rock fragments in the typical till (fig. 2) are shales, limestones (including dolomitic limestones), and dolomites. "Dolomite" in this study is rock "containing either a theoretically normal proportion or up to 10 percent excess $CaCO_3$ in the molecule" (Ramsden, 1954, p. 282). (Calcitic dolomite[3] and dolomitic limestone fragments are both lumped under the heading "limestone.") Shale fragments in the pebble and granule sizes are almost exclusively black particles of Devonian age (containing characteristic spore-like bodies), such as constitute the bedrock surface northeast of Sandusky, Ohio (fig. 3). They also form most of the shale particles in the sand sizes. In the finer sand sizes, however, a few reddish, light-gray, and green-colored shale fragments begin to appear. The clay mineral content of the tills (fig. 2) indicates that most of the shales (Ordovician to middle Devonian shales) have been ground down into component minerals that are distributed throughout the silt- and clay-size fractions of the till. Limestone and dolomite rock fragments in the till correspond with bedrock materials of this type that occur in northeastern Indiana and northwestern Ohio

[3] Mistakenly designated "dolomitic limestone" in the author's technique paper (Harrison, 1959, p. 13 and 16).

(fig. 3). Figure 4 shows the approximate weight contribution made by the important rock and mineral types to the typical till. The primary mode for the dolomite curve (fig. 4) lies in the coarsest silt size. This mode may reflect the influence of dolomite bedrock from distant sources (Dreimanis and Terasmae, 1958, p. 125), like the dolomites in the vicinity of southwestern Ontario (fig. 3) and (or) Ottawa. In this case, the secondary dolomite mode, which is in the granule size, would reflect the influence of the dolomite bedrock in Indiana and Ohio (fig. 3).

Finally, some of the rare and unusual rock types in the till find counterparts in the bedrock types of the Chibougamau Lake area (fig. 3). Rocks of the Chibougamau area have been described by Mawdsley and Norman (1935, p. 11–50).

In view of the fact that the bedrock types along the proposed flow path (fig. 3) can be considered to correspond with the rock fragments in the till and, conversely, that the rock types in the till do not suggest a different flow path than the one proposed, the approximate flow path of figure 3 is considered "verified."

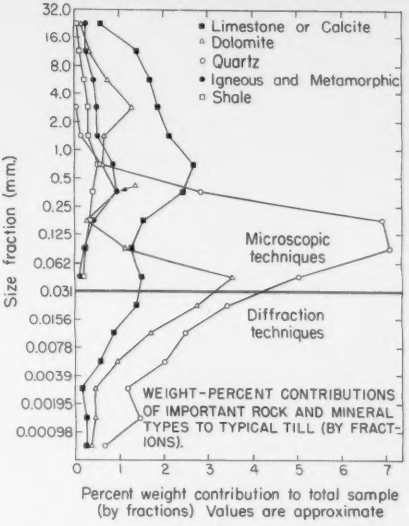


FIG. 4.—Weight-percent contributions of important rock and mineral types to typical till (by fractions).

### POSTULATED FLOW UNIT

The actual "stream"[4] of ice that flowed into Marion County is assumed to have been roughly 20 miles in width throughout its entire course (fig. 3) and is assumed to have been positioned over the lowest parts of the bedrock surface. Such precise dimensions and positioning are undoubtedly an oversimplification, but idealization of the situation is advisable for the sake of calculations. There is reason to suspect that the flow unit would have been compressed in those regions where the 750-foot contour closes together slightly (fig. 3) and that it would have expanded a bit in the broad lowland near the ice margin. It is also probable that in the upstream areas lateral flow of ice toward the streamline of most rapid movement would have carried in bedrock particles from areas outside of the 20-mile wide flow unit. The bedrock surface between Lake Huron and the western end of Lake Erie (fig. 3) lies at an altitude of less than 750 feet (Horberg and Anderson, 1956, fig. 1), but it is not likely (p. 113) that an ice stream flowing southward through this lowland would have contributed to the englacial debris of the Ontario-Erie flow unit.

It is necessary to assume a starting point and a stopping point for erosion by the ice stream (or "flow unit") of figure 3. The starting point is assumed to have been close to latitude 53° N. and longitude 71° W.; that is, a little more than 100 miles southwest of the "inferred and approximate" major ice divide in Quebec for the last major ice movement (Geol. Assoc. Canada, 1958, Glacial Map of Canada). This is believed to be a reasonable approximation for the point where the Wisconsin ice began extensive erosion of the Canadian bedrock.

The stopping point for glacial erosion is thought to have been the point where the basal ice ceased to move; that is, in the vicinity of the Wisconsin drift boundary

[4] Flow units and ice streams have been described from the Greenland and Antarctic ice sheets (Horberg and Anderson, 1956, p. 113). Mellor (1959, p. 377) has noted fast-moving ice streams from Antarctica that are related to bedrock topography and that range from 2 to 15 km in width.

TABLE 2.—*Conversion of mineral percentages to bedrock percentages*

| | Rock or Mineral Type in Typical Marion County Till | Approx. Proportion in Typical Till (Weight Percent) | Approx. Area of Bedrock Surface Passed Over (Percent) | Difference of Columns 2 and 3 (Percent) | Percentages of Mineral Types in Till Balanced | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | | | | | Quartz to Shale | Calcite to Shale | Feldspar and Dolomite to Shale | Illite to Shale | Chlorite to Shale |
| | (1) | (2) | (3) | (4) | (5) | (6) | (7) | (8) | (9) |
| Rock Percentages | Limestone | 15.0 | 18.0 | −3.0 | | | | | |
| | Dolomite | 6.0 | 17.0 | −11.0 | | | | | |
| | Shale: Black middle Devonian | 3.0 | 9.0 | −6.0 | +1.5 to 2.7 | +0.4 to 0.8 | +0.2 to 0.4 | +1.3 to 2.3 | +0.4 to 0 |
| | Shale: Ordovician | | 6.0<br>15.0 | −6.0<br>−15.0 | +1.1 to 1.9<br>+5.7 to 10.7 | +1.1 to 1.9<br>+1.6 to 3.0 | +0.5 to 0.9 | +1.5 to 2.7<br>+1.7 to 2.9 | +0.6 to 1<br>+0.5 to 0 |
| | Sandst. & Silts. (incl. graywcky.) | 0.5 | 2.0 | −1.5 | | | | | |
| | Misc. 2.5: Tillite | 1.2 | 2.0 | −0.8 | | | | | |
| | Misc. 2.5: Other | 1.3 | — | −1.3 | | | | | |
| | Ingeous (Sand grains) | 2.5 | | | | | | | |
| | Acid igneous | 0.8 | 3.0 | +0.5 | | | | | |
| | Basic igneous | 0.2 | | | | | | | |
| | Metamor. and Qtzite. | 1.0 | 28.0 | −27.0 | | | | | |
| Mineral Percentages | Calcite | 3.0 to 5.0 | | | | −3.1 to 5.7 | | | |
| | Dolomite | 7.0 to 13.0 | | | | | −0.2 to 0.4 | | |
| | Illite | 7.0 to 13.0 | | | | | | −4.5 to 7.9 | |
| | Chlorite | 3.5 to 5.5 | | | | | | | −1.5 to 2 |
| | Mixed layer minerals | 1.0 to 2.0 | | | | | | | |
| | Chert | 0.5 | | | | | | | |
| | Orthoclase | 1.0 | | | | | | | |
| | Dark minerals | 0.5 | | | | | | | |
| | Quartz (in silt-clay fract.) | 11.5 to 21.5 | | | −8.3 to 15.3 | | | | |
| | Feldspar (in silt-clay fract.) | 3.0 to 4.5 | | | | | −0.5 to 0.9 | | |
| | Quartz (in sand fract.) | 16.5 | | | | | | | |

(Ordovician Shale)

(fig. 3). The southern limit of significant bedrock influence on the composition of the Wisconsin tills, however, seems to be at about Indianapolis, according to a study of till-pebble and cobble lithologies of Marion County (Harrison, 1959, table 4). Anderson (1955, fig. 4) indicates that the lag in significant bedrock influence on the small-pebble sizes in a till may be as much as 30 miles downstream from the bedrock outcrops (here subdrift), and in areas of thick drift cover, like Marion County, this distance can be expected to be greater.

### INITIAL ESTIMATE OF PERCENTAGES OF BEDROCK PRESENT AS MINERALS

The percentages of each of the various rock and mineral types in the typical till (fig. 2) are listed in column 2 of table 2. The percentages of outcrop area of each of the various bedrock types passed over by the ice stream (fig. 3) are listed in column 3 of table 2. The percentages of outcrop area were computed by measuring the areas underlying the flow unit on figure 3 with a planimeter and then converting the measured areas to percentages of the total area of the flow unit. The literature was consulted for estimates of the relative percentages of the component bedrock types in those areas on figure 3 that are shown as being composed of more than one type of rock.

By subtracting the values of column 3 from those in column 2, some idea is obtained of the amount of each rock type that one might expect to find represented in the mineral components of the till (table 2, col. 4). For example, in column 4 one infers

TABLE 2.—(*Continued*)

| Against Percentages of Areas of Various Rock Types Passed Over by Ice | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Mixed Layer Minerals to Shale | Quartz to Sandstone Siltstone and Tillite | Dark Minerals to Sandstone Siltstone and Tillite | Feldspar to Sandstone Siltstone and Tillite | Chert Dolomite and Calcite to Limestone | Dolomite to Dolomite | Chert to Chert | Chlorite to Sandstone Siltstone and Miscell. | Approx. Remainder (Column 4 Plus Columns 5–17) | Adjustment of Surplus in Col. 18 (See p. 441) | Final Balancing | Final Remainder |
| (10) | (11) | (12) | (13) | (14) | (15) | (16) | (17) | (18) | (19) | (20) | (21) |
| | | | | +3.0 | | | | 0.0 | | | Approx. Zero |
| | | | | | +11.0 | | | 0.0 | | | " |
| 0.4 to 0.8 | | | | | | | | 0.0 | | | " |
| | | | | | | | | 0.0 | | | " |
| 0.5 to 0.9 | | | | | | | | 0.0 | | | " |
| | +0.8 | +0.1 | +0.2 | | | | +0.4 | 0.0 | | | " |
| | +0.3 | +0.1 | +0.1 | | | | +0.3 | 0.0 | | | " |
| | | | | | | +0.1 | | +1.4 | | | " |
| | | | | | | | | +0.5 | −25.1 | +25.1 | " |
| | | | | | | | | −27.0 | | | " |
| | | | | | | | | (Column 2 Plus Columns 5–17) | | | |
| | | | | −1.6 | | | | 0.0 | | | Approx. Zero |
| | | | | −1.0 | −11.0 | | | 0.0 | | | " |
| | | | | | | | | +3.0? | (Micas) ⟶ | −3.0 | " |
| 0.9 to 1.7 | | | | | | | −0.7 | +1.5? | | −1.5 | " |
| | | | | | | | | 0.0 | | | " |
| | | | | −0.4 | | −0.1 | | 0.0 | | | " |
| | | −0.2 | | | | | | +1.0 | | −1.0 | " |
| | −0.6 | | | | | | | +0.3 | | −0.3 | " |
| | | | | | | | | 0.0 | | | " |
| | | | −0.3 | | | | | +2.2 to 3.3 | | −3.3 | " |
| | −0.5 | | | | | | | +16.0 | | −16.0 | " |

that there might be a greater amount of metamorphic rock types than of limestone types represented in the mineral components of the till. The subtraction operation (col. 4) indicates that in all cases but two portions of the bedrock types passed over by the ice probably are represented in the mineral constituents of the till.

### COMPOSITION OF BEDROCK TYPES

In order to effect any conversion of mineral percentages in the till to percentages of bedrock areas, it is necessary, first of all, to find out the general mineral composition of each of the various rock types passed over. A study of the literature can provide the necessary information for some rock types, but laboratory estimates have to be made for the shales. Table 3 indicates the approximate composition of each of the three main types of shale passed over by the ice.

### CONVERSION OF MINERALS TO BEDROCK

The best way to "make bedrock" out of the mineral components of the typical till (table 2, col. 2) appears to be to start with the clay minerals and make them into shale bedrock. The bulk of the clay minerals could only have been derived from shale bedrock (Gravenor, 1954), and so one at least knows where these mineral components belong. When making shale bedrock, it is expedient to assume that each of the shale formations passed over by the ice was eroded to the same degree. This probably is a reasonable assumption, and by utilizing it a good start at making bedrock is effected.

TABLE 3.—*Composition (in percent) of shale (J. L. Harrison, Analyst)*

| Stratigraphic Position of Shale | Locality (Collector) | Composition (±30 percent of value shown) | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | | Quartz | Feldspar | Calcite | Dolomite | Illite | Chlorite | Kaolinite | Mixed Layer Minerals |
| Upper Devonian (Black) | Marion County till; sand grains (W. Harrison) | 35 | Trace | 10 | 5 | 30 | 10 | — | 10 |
| Middle Devonian; Hamilton Group (Moscow) | Cazenovia Creek, N. Y. (W. J. Wayne) | 25 | — | 25 | — | 35 | 15 | — | — |
| Upper Ordovician (Queenston) | Lewiston, N. Y. (E. R. Cumings) | 55 | 5 | 15 | — | 15 | 5 | — | 5 |

Approximate mineral values (±30 percent) will have to be used in this operation, owing to the approximate nature of the diffraction data for the minerals in the fine fractions. (Note the ranges given in columns 5–10, table 2.) A typical calculation is as follows (all column references are to table 2):

Nine percent of the bedrock surface passed over by the ice (col. 3) consists of black shale. Three percent of black shale is present as rock fragments (col. 2) in the typical till. By assuming that the shales were eroded equally, one can postulate that 6 percent of shale (col. 4) will be found in the mineral components of the typical till. This value of 6 percent may be subdivided according to the compositional percentages given for black shale in table 3. For example, 35 percent (±30 percent) of a total of 6 percent of black shale is quartz; this is written as a range: "1.5 to 2.7" percent (col. 5). This range of values is prefixed by a plus sign (col. 5) to indicate that it is being added to the *rock* percentages above the line. Percentages of quartz for the other two shale types also are listed above the heavy line in column 5. These values for the quartz content of the three types of shale bedrock are summed, and their total is subtracted (col. 5) from the values given in column 2 for quartz ("in the silt-clay fractions") *below the line*. Most of the fine-sized quartz undoubtedly was derived from the shales to begin with and does not represent crushed quartz fragments of originally larger size. Järnefors (1952, p. 191) has pointed out that even in far-travelled tills "when the size of the particles is less than about 100–50 μ, the rate of grinding seems to have been retarded." He was referring (p. 186) to tills derived from granite and gneiss bedrock.

Columns 5–10, then, give the figures for the complete process of making shale bedrock out of the approximate percentages of clay minerals, quartz, feldspar, calcite, and dolomite (all in the fine fractions). One notes that by the time that all the clay mineral percentages have been converted to areal percentages of shale bedrock, very little in the way of percentages of mixed-layer minerals or calcite is left. An important preliminary conclusion is: *the shale-making operation indicates an approximate 1 to 1 relationship between the areal percentage of shale bedrock passed over by the ice and the weight percentage of shale in the typical till.*

Does this relationship hold also for the other rock types? Let the conversion operation proceed on the assumption that it does, in order to see wherein differences between various rock types may lie. In columns 11 through 17, percentages of minerals in the till are converted to surplus areas (col. 4) of the other rock types passed over by the ice except the areas of igneous and metamorphic rocks.

### Analysis of Residual Percentages

The results of this second conversion are shown in column 17. All of the surplus areas of sedimentary rock types have been almost exactly compensated for by remaining mineral percentages. Enough quartz, feldspar, dark minerals, mica ("illite"), and chlorite remain (col. 17), however, to make up for most of the remaining 25.1 percent (col. 20) of outcrop area of metamorphic rocks passed over by the ice. Added to the "−27.0 percent" (cols. 18 and 19) to effect the final

balancing are rock fragments amounting to 1.4 percent of miscellaneous particles (of probable metamorphic derivation) and 0.5 percent of igneous rock fragments that may in reality be misidentified orthogneiss particles (see p. 435). There appears to be a slight surplus of quartz in the sand sizes. This surplus may be a result of a low estimate of the amount of sandstone and "siltstone" bedrock passed over by the ice. Conspicuous lack of rounding of quartz grains in the sand sizes, on the other hand, agrees with Anderson's observations (1957, p. 1427) for Wisconsin tills in northeastern Indiana. The author concurs that "most of these grains were derived [directly] from the disintegration of primary rocks" (1957, p. 1427), and feels justified, therefore, in balancing the quartz sand component against the surplus area of igneous and metamorphic bedrock (col. 20).

### End Result

The final remainder, after the two conversions and final balancing, is approximately zero (col. 21). The pie diagram of figure 5A shows the approximate bedrock composition of the typical till in weight percent. If one assumes for the moment the validity of the conversion of minerals into bedrock (table 2), it is possible to say that about 90 percent (by weight) of the bedrock that contributed to the typical till (fig. 2) was derived outside of Indiana. Conversely, about 10 percent of the typical till is composed of "local" (within 100 miles) bedrock materials.

### By-product

The "by-product" of the process of conversion of mineral types in the till to bedrock types passed over by the ice is an approximate relationship between the weight of a given rock type in the till and the outcrop area of that bedrock type passed over by the ice. The conversion operation (table 2) that proceeded on the assumption of a 1 to 1 relationship resulted in insignificant remainders of bedrock areas (and mineral types). It seems reasonable, therefore, to suggest that: *the relationship between the weight percent of a given rock type in the typical till and the percent of outcrop area of that rock type upstream may be about 1 to 1.*

## EVALUATION OF "END RESULT" AND "BY-PRODUCT"

### Problem of Earlier Drift

A cover of glacial drift must have existed along the flow path of figure 3 prior to advance of the Wisconsin ice. The thickness of this pre-Wisconsin drift blanket can only be estimated, but regardless of its thickness it is probable that the greatest part of it was incorporated by the advancing Wisconsin ice. What effect, then, did this drift cover have upon the composition of the typical Marion County till (fig. 2) and upon the 1 to 1 relationship deduced earlier?

The drift cover prior to advance of the Wisconsin ice was, by present-day analogy, more than three-quarters till and was deposited during the previous glaciation (the Illinoian). Samples of till of probable Illinoian age from the Toronto area, although richer in local shale bedrock, are similar in composition to overlying Wisconsin tills (Dreimanis and Terasmae, 1958, p. 124). This fact indicates that Illinoian ice followed a flow path similar to that shown on figure 3, at least as far south as Toronto. Fraction analyses of tills from central Indiana that are known to be Illinoian in age have not been made. Two samples used in this study (Harrison, 1959, nos. 11 and 13), dated by wood incorporated in till (W-814) or wood from an overlying silt (W-578) at >37,000 radiocarbon years, may be Illinoian in age (Harrison, in preparation). Gravenor (1954) published a study of an Illinoian till sample that was taken approximately 40 miles southwest of Marion County near Brazil, Indiana. The various percentages of sand and gravel, silt, and clay in this sample are almost identical to those of the typical till from Marion County (Harrison, 1959, fig. 4) The heavy-mineral suite of the Illinoian till (Gravenor, 1954, table 1, bottom 3 samples) is also very similar to that of the typical Marion County till, and it shows nearly the same ratio of purple to red garnets that exists in the typical Wisconsin till of Marion County. There are more opaque minerals and zircons in the suite of the Illinoian till (Gravenor, 1954, table 1) because of the influence of Mississippian and Pennsylvanian rocks upstream from Brazil (Gault, 1939). On the basis of these observed sim-

ilarities between the Wisconsin and Illinoian tills near the flow path of figure 3, it can be assumed that both the Illinoian and the Wisconsin tills were deposited by ice that flowed along the general path shown on this figure.

The problem to be treated is closely analogous to the problem of the readvance of a glacier over a previously deposited ground moraine. If, prior to the readvance, the till of the ground-moraine sediments had been sorted in places (as in lake basins), it makes little difference. The mixing action of the readvancing ice would "reconstitute" the sorted components into new till of nearly the same texture and composition as the previous one. Weathering of the older drift prior to a second advance would effect relatively minor compositional changes (Gravenor, 1954) and except for the surficial zone leached of carbonates the total volume of drift affected by weathering would be negligible. Textural and compositional changes prior to a readvance thus can be disregarded in the following analysis.

Imagine a cover of Illinoian drift extending all along the flow path of figure 3. The till north of Ottawa would be composed of ground up igneous and metamorphic rocks; the till at Hamilton, Ontario, would be composed of igneous, metamorphic, and sedimentary rocks (shale, dolomite, limestone; fig. 3), and the till farther south would vary according to the bedrock types encountered there. The composition of the till at any point should reflect, in some way, *all* the bedrock types upstream (Gravenor, 1951).

Now what if another ice stream were to advance over this ground moraine? It is believed that the advancing ice, by picking up the previous till, would simply be carrying on where the last ice left off. After sufficient ice had flowed over the earlier till and had stripped it from the bedrock, it is thought that the younger ice would merely be making new, but in no way different, till from the bedrock materials.

Such a process can explain the great similarity (Harrison, 1959) of the oldest and youngest tills of Wisconsin age in Marion County (fig. 1). The oldest Wisconsin tills must have been deposited by ice that incorporated both earlier Illinoian drift and fresh bedrock materials. The later Wisconsin tills, however, were derived from the earliest Wisconsin tills and from bedrock.[5]

It seems reasonable to assume that the earlier Illinoian drift cover did not have a significant effect upon the composition or texture of the typical Marion County till and that, conversely, the rough 1 to 1 relationship found earlier (and relating to *bedrock* composition) need not be discounted.

## Problem of Glacial Transport and Deposition

That the bulk of a till may consist of materials derived from relatively great distances upstream has been noted by Horberg (1953, p. 38) and by Shepps (1953, p. 47). It has been suggested here that about 90 percent of certain central Indiana Wisconsin tills have been derived from rocks between 100 and 1250 miles upstream from the area of deposition. If this be true, then what can be inferred about the mechanics of glacial transport and deposition?

One implication seems to be that materials were eroded as far away as 1250 miles and were carried in the ice all the way to central Indiana. No "plastering-on" of the till, as the ice flowed along, is indicated. All the till appears to have been carried along until the ice could flow no farther, and then it slowly melted out of the stagnant basal ice. Although the "lodgement" idea of till deposition is a popular one (Flint, 1957, p. 160), the present author finds it difficult to explain how the advancing ice stream of figure 3 could continually lodge till beneath its moving base (that is, deposit it) and, at the same time, continually acquire fragments from the local bedrock surface by

[5] If early (>38,000 $C^{14}$ years) Wisconsin tills are truly present in Marion County (cf. Harrison, in preparation), then it is probable that the entire flow path of figure 3 was deglaciated between the time of deposition of the early Wisconsin tills and the deposition of the overlying classical (<23,000 $C^{14}$ years) Wisconsin tills. With regard to the classical Wisconsin tills, the ice margin retreated only a few tens of miles between successive readvances over Marion County, and so all the younger tills represent essentially material incorporated in the ice stream during its main advance. Such a mechanism may further aid in explaining the great similarity (Harrison, 1959) of the three to five Wisconsin tills in Marion County.

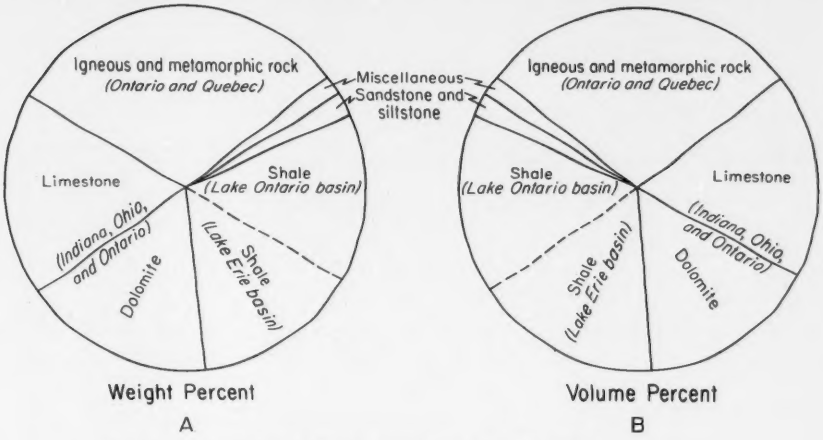


FIG. 5.—Approximate bedrock composition of typical Marion County till.

erosion (as is indicated by the composition of the typical till). The typical Wisconsin till of Marion County, therefore, is believed to have been deposited in a manner like unto that suggested for the tills of the Chicago area; that is, "by slow melting out from the stagnant basal zones of the various glaciers which transported it" (Harrison, 1957, p. 275).

### Problem of Glacial Erosion

If the 1 to 1 relationship described above is valid, then it seems reasonable to look for evidence for this relationship in terms of the relative degrees of erosion of the different bedrock formations along the flow path (fig. 3). The rocks of the Lake Ontario and Lake Erie basins, for example, have been eroded by ice to a greater degree than the bedrock of the Canadian Shield. Is this in harmony with the 1 to 1 relationship found earlier?

To answer this question it is necessary to understand that the 1 to 1 relationship found above is a relationship between the *weight* of a given rock type in the till and the outcrop area of that rock type upstream; that is, it is a weight-area relationship, not a volume-area relationship. For making estimates of the *volumes* of different bedrock materials in the typical till it is necessary to take into account the relative densities of the rock types present. The following bulk-density values (Birch and others, 1942, tables 2–1, 2–2, 2–6) have been used for conversion of weight-percent values of different rock types to volume-percent values: shale-2.3, igneous and metamorphic-2.7, limestone-2.5, dolomite-2.7, sandstone and siltstone-2.4, and miscellaneous-2.6. The conversion to volume percentages indicates that the typical till is composed of 33 percent shale, 29 percent metamorphic and igneous rock, 18 percent limestone, 16 percent dolomite, 2 percent sandstone and siltstone, and 2 percent miscellaneous rocks by volume. The pie diagram of figure 5 B shows the approximate bedrock composition of the typical till in volume percent.

On the basis of the approximate volumes of bedrock materials in the typical till, it is clear that, relatively speaking, a greater quantity of shale bedrock was eroded by the ice stream that developed the Marion County till than of the other bedrock types. This fact is in harmony with the observed overdeepening (Hough, 1958, figs. 6, 7) of the shale bedrock lowlands that presently contain the Lake Ontario and Lake Erie basins. The overdeepening of these shale lowlands, therefore, is not thought to be a result of a particularly greater "weakness" of the shale to resist glacial erosion. Intensive glacial erosion of the Devonian formations in

a small area at the east end of Lake Erie may be a reflection of two factors: (1) the greater southerly dip of the beds (Carman, 1946, p. 279) there, which permitted greater quarrying action of the ice (Zumberge, 1955, fig. 3) and (2) the ability of a flow unit of ice to depart temporarily from a simple weight rule when eroding bedrock.

Although the weight and volume percent differences of figure 5 are not large, the inference about glacial erosion that may be drawn is that the ice stream that flowed into Marion County incorporated material according to a weight rule, rather than a volume rule; that is, *the ice stream that developed the Marion County till may have eroded an equal weight—but not necessarily an equal volume—of material per unit area of surface passed over.*

An alternative to the hypothesis of excessive glacial erosion of a preglacial lowland because of the composition of its bedrock surface is the hypothesis of development of local centers of glaciation in the depressions of this preglacial lowland. Dreimanis (1951) suggested a local center of glaciation in the Lake Erie basin, and both Holmes (1952) and Dreimanis and Terasmae (1958, p. 134) found evidence of a local center of glaciation in the Lake Ontario basin. Overdeepening of the lake basins may possibly have been restricted largely to the latter phases of the Wisconsin (and earlier?) glaciations. It seems likely, however, that both methods of overdeepening were operative.

## TESTING THE RELATIONSHIP

Admittedly, the 1 to 1 relationship found earlier is sensitive to the positioning and width of the flow path and to the "assumed starting point of glacial erosion" (fig. 3). The relationship is believed to be of such potential importance to a closer understanding of the mechanics of glacial erosion, transportation, and deposition, however, that it warrants further investigation as to its validity. The relationship could be tested by collecting samples of basal Wisconsin till at strategic points along the flow unit of figure 3 and subjecting them to fraction analysis (following Harrison, 1959, so that results could be compared). Not only could this relationship be proved or disproved, but also many basic data relative to the abrasional characteristics of the various bedrock types along the flow path might be determined.

## APPLICATION OF STUDY METHODS TO PROBLEMS OF GLACIAL HISTORY

Applications of the methods and results of this study in the field of glacial sedimentation have already been mentioned. The greatest value of studies of till composition will undoubtedly come, however, in the interpretations which they afford of glacial history. A case in point is the till associated with the Shelbyville drift in Illinois (Geological Society of America, 1959, Glacial Map of North America East of the Rocky Mountains). Was it developed by ice streams flowing along the Ontario-Erie lowland, or by ice of the Huron-Saginaw (Horberg and Anderson, 1956, table 2) or Lake Michigan lobes? Fraction analysis of the Shelbyville till, coupled with an analysis of till provenance like that of the present study, might provide the answer. It would be expected, for example, that if the Shelbyville till were developed by an ice stream that flowed along the Ontario-Erie lowland, it would differ in composition from the typical till in Marion County (fig. 2) only by the amounts of the different bedrock materials that would have been incorporated as the ice stream flowed through north-central Indiana and east-central Illinois. By determining the provenance of the Shelbyville till, information as to the influence of a specific center of glaciation on the deposition of this drift could be obtained, as well as a clearer understanding of the sequence of glacial lobes in this area of the Central Lowland.

## SUMMARY AND CONCLUSIONS

The original bedrock composition of till of Wisconsin (and Illinoian?) age from Marion County, central Indiana, was approximately determined according to the following steps:

1. The texture and composition of a "typical" till, based on averaged data from 11 samples (fig. 2), was determined in detail by means of fraction analyses (Harrison, 1959).

2. The flow path followed by the ice streams that developed the typical till was postulated and then was verified (approximately) by comparisons of bedrock frag-

ments and heavy minerals in the till with bedrock types along the flow path. A flow *unit* 20 miles wide and centered over the lows in the bedrock surface then was postulated as having the precise dimensions of the ice stream that moved along this flow path into Marion County.

3. The mineral composition of the various bedrock types passed over by the flow unit was assessed, either by literature studies or by laboratory analyses.

4. Mineral components of the typical till were converted into bedrock types passed over by the ice. The conversion process (table 2) began with the translation of weight-percent values for clay minerals into areal percentages of shale bedrock passed over by the ice. The conversion of clay and associated minerals into shale bedrock was accomplished on the assumption that each type of shale was eroded to the same extent and also was represented in the typical till in an amount proportional to its area of outcrop upstream. This "successful" conversion prompted an attempt to convert the remaining mineral components into the other bedrock types by following, as a first attempt, the assumptions used when making shale bedrock. This last attempt, within the framework of the study, was also successful.

Results of the foregoing analysis indicate that in terms of weight-percent composition the typical till (fig. 2) is composed of about 30 percent shale, 28.5 percent metamorphic rock, 18 percent limestone, 17 percent dolomite, 2.5 percent igneous rock, 2 percent sandstone and siltstone, and 2 percent other rocks (fig. 5A). The high weight-percentage of shale in the typical till does not indicate a lesser resistance to erosion of the shale bedrock; when bulk densities of the different bedrock types are considered, however, it is clear that, relatively speaking, a greater *volume* of shale was eroded than any other rock type. In terms of volume-percent, the typical till is composed of approximately 33 percent shale, 29 percent metamorphic and igneous rock, 18 percent limestone, 16 percent dolomite, 2 percent sandstone and siltstone, and 2 percent miscellaneous rocks (fig. 5B). Percentages of bedrock types in the typical till indicate that about 90 percent of it (by weight or volume) consists of bedrock materials derived from outcrops more than 100 miles upstream from the locality of till deposition.

Enough minerals and (or) rock fragments were present in the typical till to "make bedrock" in amounts equal to the percentages of area of each of the bedrock types traversed by the ice (table 2). The ratio between the weight-percent of a given rock type in the typical till (fig. 2) and the percent of outcrop area of that rock type passed over by the ice, therefore, seems to be about 1 to 1. This relationship is sensitive to the positioning and width of the assumed flow path and to the assumed starting point of glacial erosion.

Investigation of the validity of the 1 to 1 relationship is desirable because of its potential value to an understanding of the processes of glacial sedimentation. If it is true, for instance, it indicates that the Marion County tills were not deposited by melting out from (or lodgement beneath) the bases of moving ice streams and that bedrock materials may be picked up and transported by glacial ice for distances in excess of 1200 miles. The relationship also indicates that the bedrock was eroded according to a weight rule.

ACKNOWLEDGMENTS

The author is indebted to the following workers for critical review of the manuscript: L. A. Bayrock, A. Dreimanis, P. E. Potter, and V. C. Shepps. G. W. White kindly commented on an early draft of the paper, and A. W. Forslev provided discussion. The author is solely responsible for the conclusions presented and for the reasoning employed in reaching them. Radiocarbon age determinations were made in the Washington Laboratory of the U. S. Geological Survey.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 447–458
FIGS. 1–10, SEPTEMBER, 1960

# CROSS-LAMINATION ANALYSIS IN THE UPPER CAMBRIAN FRANCONIA FORMATION OF WISCONSIN[1]

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## ABSTRACT

The type of cross-lamination most commonly observed in the Upper Cambrian Franconia Formation of southwestern Wisconsin is trough cross-lamination. Statistical analyses of 1023 cross-lamination dip azimuths in the sandstone members of the Franconia Formation indicate a preferred orientation of sedimentary transport 151 degrees to the southeast. The mean dip of the individual cross-laminae in 700 measurements is 18 degrees. A comparison of sand size of individual cross-laminae with degree of dip indicates that an increase in dip is associated with good sorting and coarser grain size.

The Precambrian upland to the north of the area (or strata in equivalent position) is believed to have supplied sediments to the Franconia seas. Cross-lamination (azimuth) study of the Precambrian Baraboo Quartzite within the area also reflects this northerly source. Similar evidence obtained from regional studies of Huronian and Upper Cambrian sediments indicates the possibility of a stable regional paleoslope throughout a substantial period of geologic time.

## INTRODUCTION

Recent analyses of cross-lamination have proven of considerable value in determining the paleogeography and sedimentary regional-slope of sandstone formations (Potter and Siever, 1956, p. 448; McDowell, 1957; Pettijohn, 1957; Rusnak, 1957; Hamblin, 1958; Pelletier, 1958). In many of these studies fluvial or deltaic sedimentation appears to have been dominant. Few investigations of cross-lamination in a shallow marine deposit such as the Franconia Formation have been made.

The purpose of this investigation is to study the cross-lamination in Upper Cambrian Franconia Formation of southwestern Wisconsin. The Mazomanie-Reno Members of the Franconia Formation were chosen for the following reasons: (1) the abundance of cross-laminated units, (2) proximity to the strand line as indicated by previous investigators (Berg, 1954; Twenhofel, Raasch, and Thwaites, 1935), (3) to test the thesis that cross-lamination developed in a formation of unquestionable shallow marine origin could be used to determine the paleoslope.

## THE FRANCONIA AND ASSOCIATED FORMATIONS

The Upper Cambrian Franconia Formation in west-central Wisconsin is a glauconitic, well sorted quartzose sandstone that varies from 100 to 160 ft in thickness. The formation is bounded at the top by the dolomitic strata of the St. Lawrence Member of the Trempealeau Formation and at the bottom by the white, medium-grained sandstone of the Galesville Member of the Dresbach Formation. The basal greensand conglomerate, considered until recently to be the base of the Trempealeau Formation, has been redefined by Berg, Nelson, and Bell (1956, p. 8) to be the upper limit of the Franconia Formation.

The relationship of the Franconia Formation to the other formations within the Croixian Series and to the other series within the Cambrian System of southwestern Wisconsin is noted in figure 1. The members of the Franconia Formation are based on rock types designated by Berg (1954, p. 857):

"They are, in ascending order: the Woodhill member—medium to coarse-grained sandstone; the Birkmose member—fine-grained glauconitic sandstone; the Tomah member—sandstone and shale; and the Reno member—glauconitic sandstone. A fifth unit, the Mazomanie member—thin-bedded or cross-bedded sandstone, forms a nonglauconitic facies that interfingers with and replaces the Reno member. Faunal zones are independent of the lithic units.

The Woodhill member was deposited in a transgressing Franconia sea. Birkmose greensands formed in shallow waters far from shore, while Tomah sand and shale and Mazomanie

[1] Manuscript received November 9, 1959.

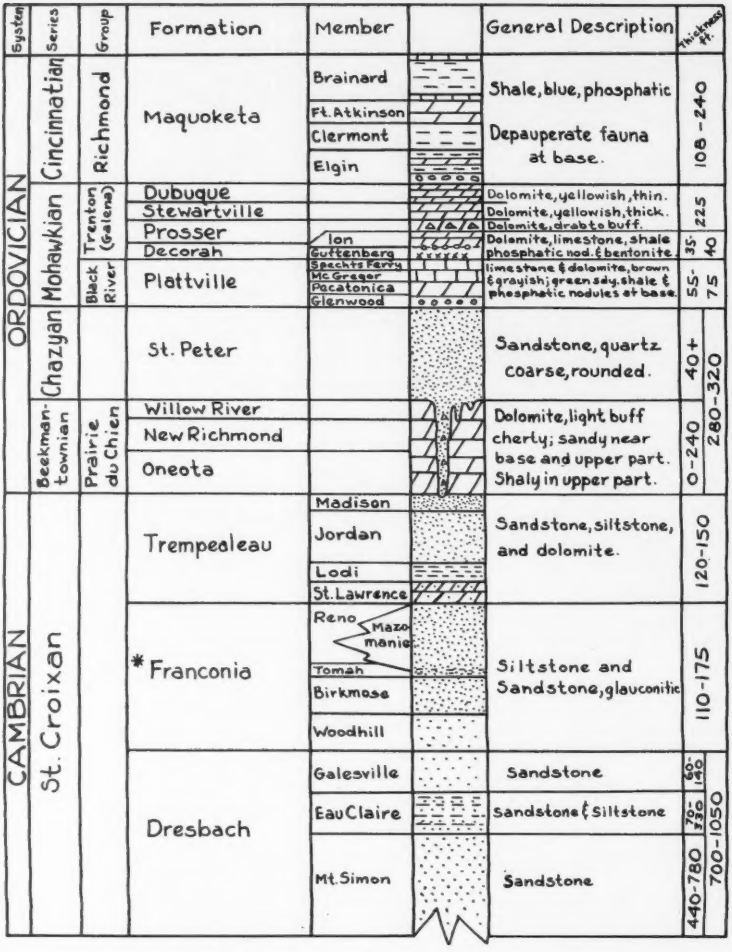



FIG. 1.—Generalized stratigraphic column for southwestern Wisconsin. After figure 1, Conference Classification, Kansas Geol. Survey, 9th Annual Field Conference, 1935. Modified after Agnew, A. F., and others, 1956, p. 255.

* Franconia Formation from Berg, 1954, p. 858.

thin-bedded sand was deposited shoreward, and cross-bedded Mazomanie sand accumulated nearest the strand line."

The dominant outcrops in the study area are of the Reno and Mazomanie Members. It was not possible to differentiate in the field between the Reno and Mazomanie Members as defined by Berg. Bands of sandstone containing high percentages of glauconite are found interspersed throughout most

of the Mazomanie Member, whereas the Reno Member contains many sandstone beds with little or no glauconite.

Cross-laminated strata are common throughout most of the Franconia Formation. The thinly bedded, fine-grained micaceous sandstones and shales of the Tomah Member are not cross-laminated. Cross-lamination was noted in the Galesville Member of the Dresbach Formation and in the Jordan Member of the Trempealeau Formation, which lie below and above the Franconia Formation, respectively.

## HISTORY OF THE FRANCONIA NOMENCLATURE

An excellent summary of the nonemclature history can be found in a paper by Trowbridge and Atwater (1934, p. 45–50) and more recently by Berg (1954, p. 858–859).

Twenhofel, Raasch, and Thwaites (1935, p. 1703) subdivided the Franconia Formation into several faunal zones. The basal conglomerate and greensand which occurs above the Franconia Formation was considered by Twenhofel, Raasch, and Thwaites (1935, p. 1705) to be a part of the St. Lawrence Member of the Trempealeau Formation. Berg, Nelson, and Bell (1956, p. 8) recognized that faunal units in the Franconia Formation climb shoreward with respect to lithic changes. They concluded (1) that the Franconia-Trempealeau boundary is conformable, (2) there is no faunal stage that coincides with the top of the Franconia Formation, and (3) the basal greensand conglomerate of the Trempealeau Formation would be more suitably placed with the similar lithology, characteristic of the Franconia Formation.

Snodgrass (1955) and Hass (1955) concluded that faunal zones in the Franconia Formation of southwestern Wisconsin were difficult if not impossible to recognize in the field. Ericson (1951) also recognized the need for better methods than faunal zones to subdivide the Franconia Formation into easily discernible units. Berg (1954, p. 857–882) recognized this problem and redefined the Franconia Formation in terms of lithic units. These units have already been mentioned as members of the Franconia Formation and will be used throughout this paper.

## FIELD METHODS

A total of 1023 dip directions of cross-lamination were taken within the Franconia Formation. Additional measurements were taken within the Galesville Member of the Dresbach Formation and the Jordan Member of the Trempealeau Formation; these measurements, totaled 30 and 45 respectively.

Cross-lamination inclinations and dip azimuths were measured by means of a Brunton compass. Thickness measurements were made with a 6-in steel rule. Adjustment of cross-lamination inclinations to compensate for regional dip was not considered necessary (Tanner, 1955, p. 2472) because the regional dips of the Franconia and associated formations are usually less than 2 degrees in Wisconsin.

Thickness measurements were made of the cross-laminae within each cross-laminated set along with the thickness of the cross-laminated set or bed. One azimuth measurement was made for each cross-laminated unit. Care was taken to analyze each cross-laminated unit and to take the true direction of inclination of the unit along with the maximum or true dip reading (fig. 2). Concentration of glauconite along bedding planes and differential weathering of the Mazomanie Member of the Franconia Formation caused many of the cross-laminae to stand out in bold relief. From 2 to 25 cross-lamination measurements were made at each outcrop.

The Township-Range grid pattern was used early in the study to obtain an even distribution of samples between townships. It was soon discovered, however, that outcrops had to be taken wherever they were encountered because of their scarcity and poor exposure. These factors in turn affected an uneven sampling distribution.

Size analyses were made of 21 sand samples collected from the Franconia Formation to see if any relationship existed between sand size and angle of inclination.

## CROSS-LAMINATION

### Type

Cross-lamination of two types was observed in the Franconia formation. These types are trough and planar cross-lamination (McKee and Weir, 1953, p. 387).

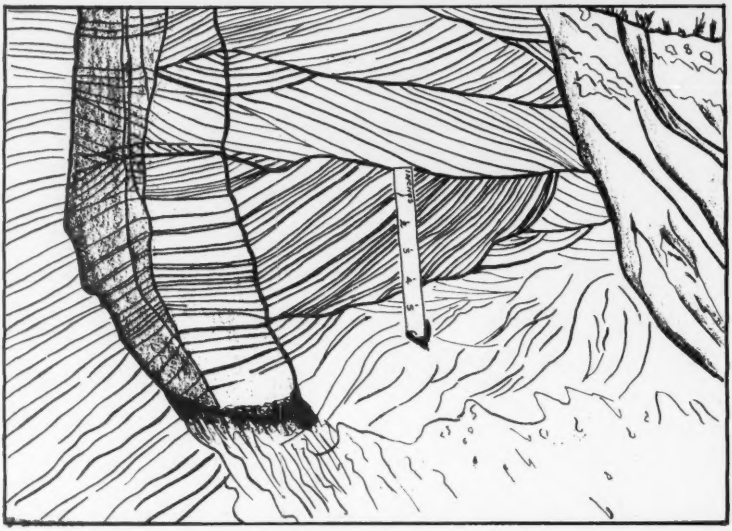


FIG. 2.—A typical example of trough cross-lamination as found in the Mazomanie-Reno Members of the Franconia Formation. Unit is exposed at a right angle to the face of the outcrop by trench on the left. Location: SW¼ SE¼ sec. 24 T. 9 N. R. 5 E., Sauk County, Wisconsin.

Trough cross-lamination is the most common, whereas planar cross-lamination was observed in only a few instances. Potter and Siever (1956, p. 231–232) attribute the variation in cross-lamination types to variation in degree or kind of turbulence at the time of deposition. Vigorous turbulence would produce trough cross-lamination, and less vigorous turbulence would favor the formation of planar cross-lamination.

Essentially no differences were observed between the types of cross-lamination in the Galesville, Jordan or Mazomanie-Reno Sandstones.

## Thickness

The thickness of the measured cross-laminated units ranges from 1 to 32 in. The average thickness is 10 in. In most cases the cross-laminated units were truncated on the upper surfaces and were followed by deposition of additional cross-laminated units or flat-lying beds. Experimental evidence indicates that if the upper surface of a cross-laminated unit is to be preserved, the water level must remain constant throughout deposition (McKee, 1957, p. 131–132). If a slight lowering of the water level is effected, the upper surface of the sedimentary unit will be raised above the base level of deposition and it will then be eroded.

In a shallow marine shelf sand, such as the Franconia, cross-laminated sedimentation unit thickness would be expected to be constantly changing because of greater opportunity for reworking of sediments prior to final burial. This variability of unit thickness would be caused by wind waves and tidal, rip, and jet currents (Potter and Siever, 1956, p. 449).

The thickness of individual cross-laminae were observed to range from platy (⅛ in) to flaggy (2 in). Concentrations of glauconite along the surfaces of most cross-laminae make them easily discernible. The greatest majority of the cross-laminae varied from ⅛ to ½ in thick.

Schwarzacher (1953, p. 325–326) states that thin units of cross-lamination are most apt to occur under conditions of fluctuating current velocity and sedimentary load. The majority of the sedimentation units and cross-laminae in the Franconia Formation are thin.

### Inclinations

The average angle of inclination of the cross-laminae in the Franconia Formation was found to be 18 degrees for a total of 725 measurements. The angles of inclination of cross-laminae range from 2 to 36 degrees. The average inclination of the cross-laminae falls well within the limits of sandstones (fig. 3) which are considered to be subaqueously deposited (Twenhofel, 1950, p. 245). The mean value of 18 degrees found for cross-laminae in the Franconia Formation closely approximates the mean values found by other investigators for cross-laminae in flat-lying and moderately tilted formations (Pelletier, 1958, table 4).

A grain size analysis of 21 samples of Franconia Sandstone indicates a definite correlation between grain size and the inclination of the cross-laminae. A comparison of two or more samples at the same outcrop suggests that finer grained sediments were deposited at lower angles of inclination than were sediments of a larger grain size. The coefficient of sorting, however, indicates a correlation between high angles of inclination and better sorting (fig. 4). McKee (1957, p. 130) gives a number of factors which might influence and affect a greater angle of repose for the laminae of a cross-laminated unit during deposition; two of these, poor sorting and larger grain size, appear as contributing factors. Among the samples measured by the author, good sorting and larger grain size appear to be related to a steep angle of repose.

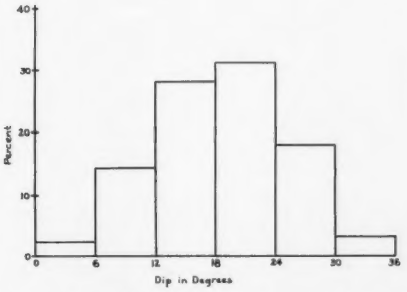



FIG. 3.—Histogram showing cross-bedding inclinations in the Franconia Formation of southwestern Wisconsin.

### CROSS-LAMINATION DIP DIRECTION

The principal objective of this paper is the study of the cross-lamination dip direction of the Mazomanie-Reno Members of the Franconia Formation. The estimated preferred orientation of the cross-lamination dip azimuths of these combined members is 165 degrees with a standard deviation of 81 degrees.

The Tukey orientation test as described by Harrison (1957, p. 101–102); Rusnak (1957, p. 53-54), and Potter and others (1958, p. 1022–1024) was used to obtain the preferred orientation of the cross-lamination azimuths. This test combines a vectorial approach with the Chi-square distribution test for statistical significance and is especially valuable in treating poorly oriented data (Potter and others, 1958, p. 1023).

In one case the Chi-square value of cross-lamination azimuths fell into the 90th percentile of significance as is indicated for two degrees of freedom in the Chi-square tables of Dixon and Massey (1957, p. 385), whereas in the remaining 6 cases, the Chi-square values fell well within the 99.5 percentile of significance. Thus there is a preferred orientation of cross-lamination azimuths within each of the formation members.

The preferred orientation of cross-lamination dip azimuths in other members within the Franconia Formation are 161 degrees for the Birkmose Member and 171 degrees for the Woodhill Member. The combined preferred orientation of cross-lamination azimuths for the Franconia Formation is 151 degrees. Preferred orientation of cross-lamination dip directions in the Galesville Member of the Dresbach Formation is 253 degrees, whereas within the Jordan Member of the Trempealeau Formation it is 199 degrees (see table 1 for statistical summary of cross-lamination azimuths).

Comparison of cross-lamination dip directions (azimuth) of other formations with those of the Franconia Formation indicates that the standard deviation of 81 degrees for the Franconia Formation is quite high (table 1). The standard deviation found for cross-lamination azimuths in the Coconino Sandstone varied between 38 and 64 degrees (Reiche, 1938, p. 915). Pettijohn (1957, p. 473) calculated the standard deviation of dip

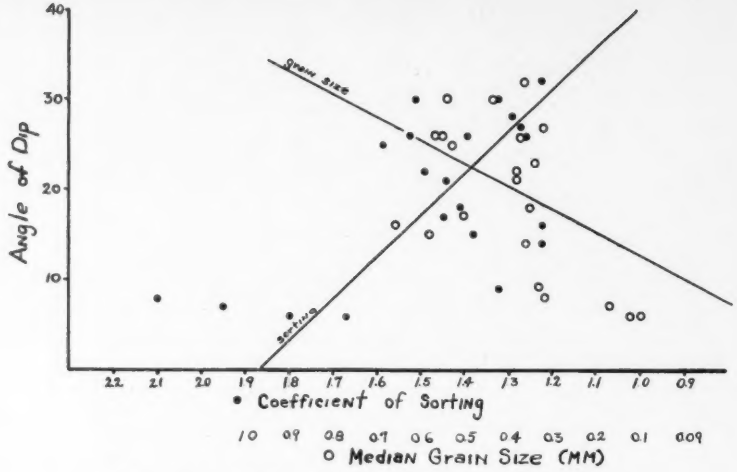

Angle of Dip
40
30
20
10
Grain Size
Sorting
22 21 20 19 18 17 1.6 1.5 1.4 1.3 1.2 1.1 1.0 0.9
• Coefficient of Sorting
1.0 0.9 0.8 0.7 0.6 0.5 0.4 0.3 0.2 0.1 0.09
○ Median Grain Size (MM)

FIG. 4.—Relationship between angle of dip and size and sorting.

directions of cross-lamination in the Lorrain Quartzite to be about 50 degrees. The standard deviation of cross-lamination dip direction of 81 degrees in the Franconia sandstones is almost twice as large as that found by Tanner (1955) to be the standard deviation of stream cross-lamination dip direction on the coastal plain of Georgia. The high standard deviation of azimuth directions in the Franconia Formation indicates factors which influenced the development of azimuth directions of the cross-laminated units

TABLE 1.—*Statistical Summary of Cross-Bedding Azimuths*

| Formation and Place | Mean | Standard Deviation | No. Readings | References |
|---|---|---|---|---|
| Franconia, Wisconsin | | | | |
| (a) Mazomanie-Reno Member | 165 | 81 | 900 | This report |
| (b) Birkmose Member | 161 | 76.5 | 31 | This report |
| (c) Woodhill Member | 171 | 50.5 | 92 | This report |
| Dresbach, Wisconsin | | | | |
| Galesville Member | 253 | 80.9 | 30 | This report |
| Trempealeau, Wisconsin | | | | |
| Jordan Member | 199 | 108 | 45 | This report |
| Pleasantview, Illinois | 357 | 67 | 30 elements | Rusnak (1957) |
| *Missisagi, Bruce Area | 100 | — | 15 | Pettijohn (1957) |
| Lorrain, Bruce Area | 163 | 58 | 119 | Pettijohn (1957) |
| Lorrain, Lacloche | 177 | — | 32 | Pettijohn (1957) |
| Mesnard, Michigan | 97.5 | — | 16 | Pettijohn (1957) |
| Sturgeon, Michigan | 143 | 38 | 80 | Trow (1948) |
| Baraboo, Wisconsin | 163 | — | 283 | Brett (1955) |
| Sioux, South Dakota, Minnesota and Iowa | 162 | — | 39 | Brett (unpub.) |
| Coconino, Arizona | — | 38 to 64 | More than 50 per locality | Reiche (1938) |
| Streams, Georgia Coastal Plain | 103 | 42 | 64 | Tanner (1957) |

* Lower part of table after Pettijohn (1957).

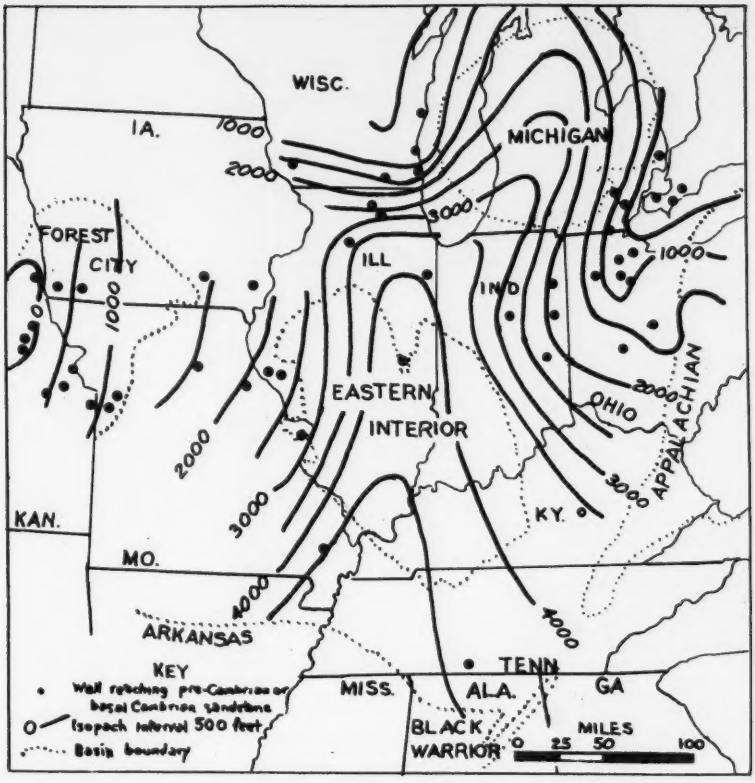



FIG. 5.—Thickness of pre-St. Peter (Lower Ordovician and Cambrian) rocks, Eastern Interior basin. Basins are outlined by sea-level contours on top of Devonian limestone. (After Swann and others, 1951, p. 496, fig. 153.)

to be considerably variable. Potter and Siever (1956, p. 449) suggest that sands deposited in a marine shelf environment might be expected to show a greater dispersion of cross-lamination azimuths when compared with azimuths of fluvial sands which are possibly less subject to working prior to final burial.

The deposition of the Franconia sandstones in a shallow marine environment is substantiated by the presence of abundant glauconite. Analysis of the authigenic feldspar content of the Tomah Member of the Franconia Formation by Berg (1952, p. 223) led him to believe that this member was deposited in a shallow marine sea. The fossil evidence and presence of mud cracks at several levels in the Franconia Formation also point out a similar shallow, perhaps littoral, marine environment (Wannemacher, Twenhofel, and Raasch, 1934, p. 20).

The mean azimuth direction of cross-lamination in the Franconia Formation is 151 degrees to the southeast. An isopach map (fig. 5) (1951) of pre-St. Peter sediments in the Eastern Interior Basin (Swann and others, 1951, p. 496) indicates by a thickening of sediments that the regional sedimentary basin was to the south and southeast

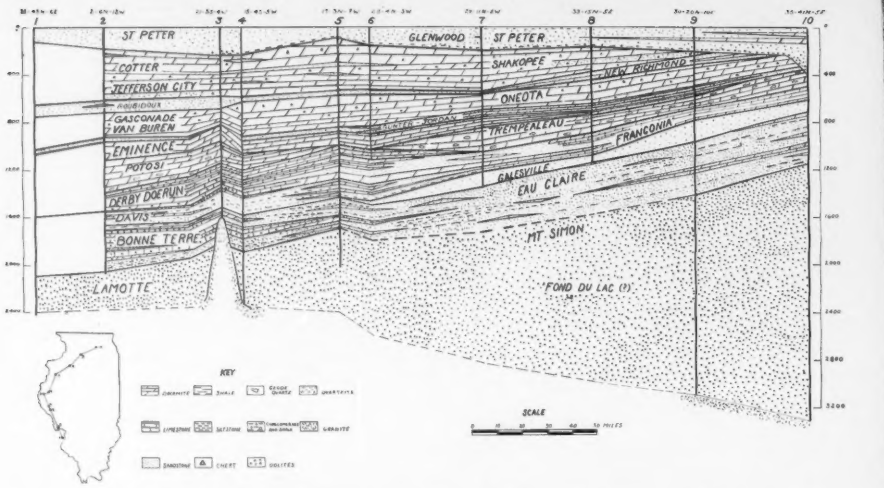



FIG. 6.—Cross-section illustrating thickening and facies change of the Franconia Formation from sand to carbonates across Illinois from north to south suggesting basinward deposition to the south. (Cross-section after Workman and Bell, 1948, p. 2047, fig. 3.)

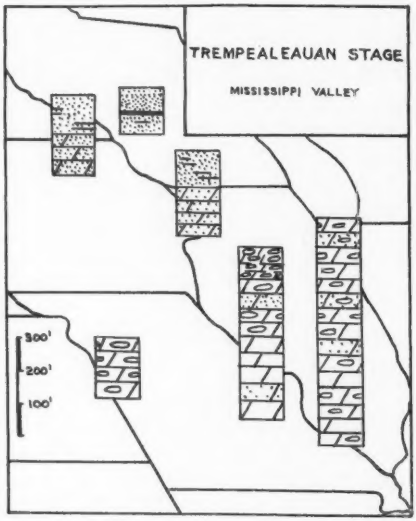



FIG. 7.—Representative sections of uppermost Cambrian strata in Mississippi Valley, illustrating gradation of time-stratigraphic unit from sandstone to dolomite in stable shelf environment. Thickening to south suggests basinward conditions. (After Dapples, Krumbein, and Sloss, 1948, p. 1928, fig. 2.)

during St. Croixian and Beekmantown time. The thinning to the north (see fig. 6) is not entirely caused by depositional effects and may be attributed in part to pre-St. Peter uplift and subsequent erosion (Sutter and others, 1959, p. 31). This pre-St. Peter unconformity is considered one of the most important breaks in the Paleozoic stratigraphy of the craton and St. Peter Sandstone is found deposited on beds that range from Shakopee to Franconia in age (Gutstadt, 1958, p. 52; Sutter and others, 1959, p. 31). Gutstadt (1958, p. 41) correlates the Franconia sandstones with finely crystalline dolomite of the lower Knox Group in Indiana. Gutstadt (1958, p. 82) suggests that Upper Cambrian sands northwest of Indiana represent beach and strand line deposits while to the southeast a contemporaneous facies of the lower Knox represents the seaward deposition of carbonate rocks. Dapples, Krumbein, and Sloss (1948, p. 1928) show several sections of the Trempealeau and equivalents from Wisconsin to Missouri (fig. 7), illustrating the change from sandstone to limestone without an intervening shaly facies. They believe this southward thickening of carbonates suggests a transition toward basinal conditions. This transition from sandstone

to limestone in a southerly direction is also characteristic of the Franconia Formation. A cross-section by Workman and Bell (1948, p. 2046) shows the Franconia expanding greatly across Illinois to the south (fig. 6). It ranges from 100–160 ft of sand in the north to a known maximum of 440 ft of sandy dolomite and limestone in the south. Thus, although a definitive answer depends on a detailed regional facies map, the present evidence suggests that the cross-lamination of the marine shelf sandstones of the Franconia Formation indicates a dominant basinward paleoslope movement of the sediments.

Although the standard deviation for the cross-lamination dip directions of the formations studied is large, the mean azimuth for each formation indicates a surprisingly consistent sedimentary source to the northwest. Heavy mineral and sand size analysis by Driscoll (1959, p. 13–14) indicates the Franconia Formation to be the transgressive phase of a transgressive-regressive sequence of sedimentation. Driscoll implies that this transgression of Franconia sandstones developed as a nearshore deposit as the Franconia sea advanced in a northerly direction over Wisconsin. Therefore, it appears that the Precambrian upland to the north of the study area or strata in equivalent position must have been the source supplying sediments to the Franconia seas.

Hamblin (1958, p. 137) notes a northerly source for the Miner's Castle Sandstone of Northern Michigan which he directly correlates with the Franconia and Trempealeau Formations of Wisconsin. Hamblin states that sediments in Northern Michigan were derived locally from the south prior to Miner's Castle time, and this southerly source is expressed in the analysis of the cross-lamination within the underlying Jacobsville Formation and the Chapel Rock Member of the Munising Formation. However, he suggests that by the advent of Miner's Castle time the Northern Michigan Highland had been essentially eroded and the regional sedimentary slope established to the southwest.

Work by Hoffman (1953) on sand size analysis of the Cambrian formations of Wisconsin indicates a possible source to the northeast for sediments of Upper Cambrian age in Wisconsin. Hoffman noted that finer sediments occur on the southwest sides of known Precambrian highs during Upper Cambrian time. He concluded that these Precambrian highs provided a sheltered area from prevailing northeasterly Upper Cambrian wind currents, whereas in quieter water the finer sediments could be deposited. Further evidence that sediments may have been derived from a northerly direction is noticeable upon examination of the outcrop average cross-lamination dip directions on the map in figure 8. The dip measurements just south of the Baraboo Range have unusually large dispersion. Away from the Baraboo Range, which was a Precambrian high in Franconia time (Wannemacher, Twenhofel, and Raasch, 1934, p. 2), the cross-lamination dip directions have a more uniform distribution. The greater variability just south of the Baraboo Range may have been caused by the interruption of the prevailingly seaward depositional slope. Secondary factors such as littoral drift and tidal currents might have influenced cross-lamination dip directions to vary more in this area than what is expected for sandstones that have had a dominantly fluvial or deltaic origin.

Examination of the rose diagram "D" in figure 9 indicates a definite preferred orientation for cross-lamination azimuths to the southwest in the Mazomanie-Reno Members of the Franconia Formation. A comparison of the azimuthal measurements and mean azimuth between that of the Baraboo Quartzite, 163 degrees (Brett, 1955, p. 146), the Lorrain Quartzite, 163 degrees (Pettijohn, 1957, p. 473), and the Mazomanie-Reno Members, 165 degrees, shows substantially few differences (fig. 9). McDowell (1957, p. 30), upon detailed investigation of the Missisagi Quartzite, found the mean dip direction of cross-lamination to be 158 degrees for the Lower Missisagi Quartzite and suggests a source for these sediments to the northwest in the Central Lake Superior region. The distribution of cross-lamination azimuths in the Mazomanie-Reno Members of the Franconia Formation compares closely with that given for the Baraboo Quartzite by Brett (1955, p. 146). Directional features of the Franconia sediments indicate that the pattern of sedimentary transport was essentially similar to that of

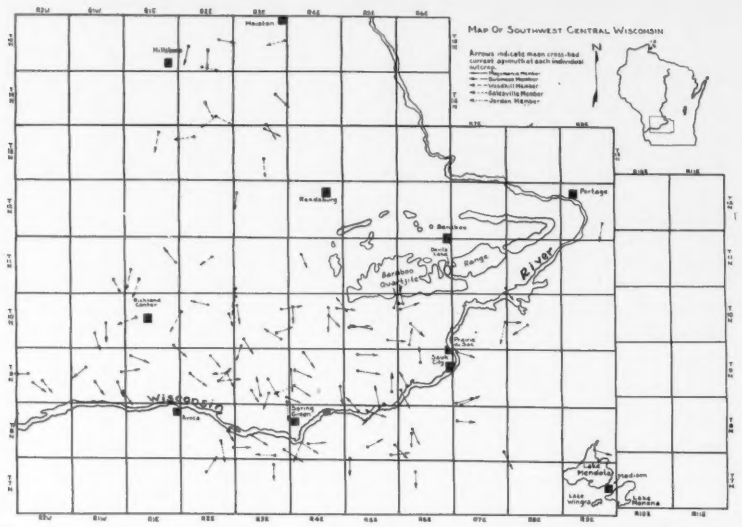
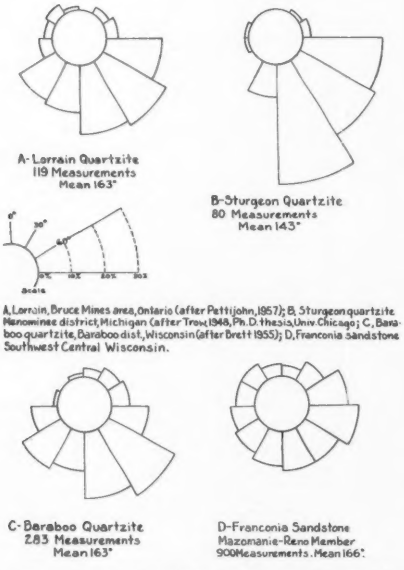



FIG. 8.—Azimuths of mean cross-bedding dips at individual outcrops in southwest-central Wisconsin.




A, Lorrain, Bruce Mines area, Ontario (after Pettijohn, 1957); B, Sturgeon quartzite, Menominee district, Michigan (after Trow, 1948, Ph. D. thesis, Univ. Chicago); C, Baraboo quartzite, Baraboo dist., Wisconsin (after Brett 1955); D, Franconia sandstone, Southwest Central Wisconsin.

FIG. 9.—Cross-bedding azimuth distribution in Lake Superior Precambrian quartzites compared with the Franconia sandstone.

the Baraboo Quartzite and other regionally located Huronian sediments even though separated by one of the largest unconformities in the Upper Mississippi Valley region.

## CONCLUSION

The great majority of cross-lamination observed in the Franconia Formation is of the trough cross-lamination type (fig. 10). The mean inclination of the cross-laminae falls well within the limits allotted to the angle of repose of a water deposited sand.

Cross-lamination in the Mazomanie-Reno Members of the Franconia Formation indicate a definite preferred azimuthal orientation to the southeast. The standard deviation at member level is found to be high. Deposition on a shallow marine shelf may be the principal reason for this. Cross-lamination measurements in other members of the Franconia Formation and members of adjacent formations also reflect a northerly source area. The Precambrian Huronian Baraboo Quartzite in the vicinity of the area studied and regionally located Huronian sediments also reflect this northerly source.

## ACKNOWLEDGMENTS

The author wishes to thank Dr. Stanley A. Tyler for guidance throughout this study.

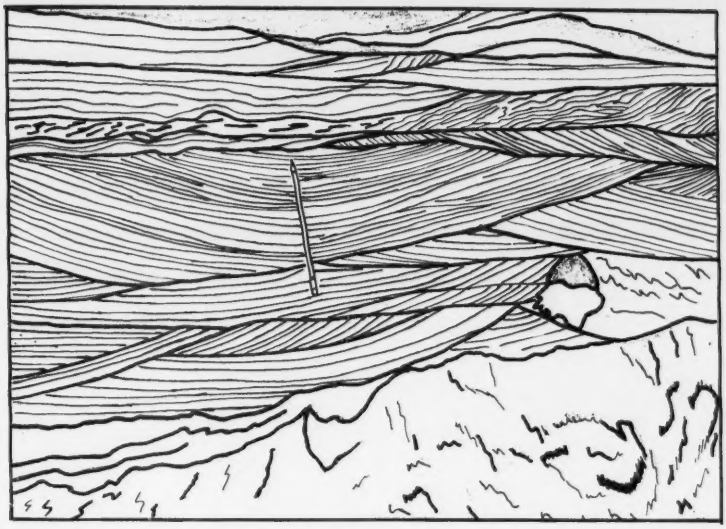


FIG. 10.—Cross-section view of a typical cross-laminated sedimentation unit within the Mazomanie-Reno Members of the Franconia Formation. Location: SW¼ SE¼ sec. 24, T. 9 N., R. 5 E., Sauk County, Wisconsin.

The writer wishes to thank Professor Frederick T. Thwaites whose aid in the field added immeasurably to the success of this problem. Acknowledgments are also due to Paul E. Potter and Wayne Pryor for helpful advice and criticism of the manuscript. Thanks also to George F. Hansen, State Geologist of Wisconsin, and Jack Steuerwald, Assistant State Geologist of Wisconsin, for their advice and numerous suggestions.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 459–465
FIGS. 1–5, SEPTEMBER, 1960

# CROSS-BEDDING DIRECTIONS IN UPPER TRIASSIC SANDSTONES OF WEST TEXAS[1]

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## ABSTRACT

Upper Triassic non-marine sandstones and shales of the Dockum Group occupy a widespread area in west Texas. The source area(s) for these beds has been a subject of varied opinion. Dip directions of foreset beds within the commonly cross-bedded sandstones were studied at 108 outcrops scattered throughout 7 counties in order to determine the direction of regional paleocurrents. Results show a predominance of dips to the northwest and west. The inferred source areas lie mainly to the southeast and east of the depositional area. The Llano Uplift region of central Texas, lying to the southeast of the area studied, was a major contributor of sediments during Late Triassic time.

## INTRODUCTION

Widespread outcrops of non-marine sediments of Late Triassic age occur in west Texas, extending from the Canadian River valley in the Texas panhandle southward to Mitchell County. The Triassic sediments present in the southern part of the outcrop area occupy much of Crosby, Garza, Borden, Scurry, and Mitchell counties (fig. 1). The beds are assigned to the Dockum Group and generally consist of thin, gray, massive or cross-bedded sandstone units alternating with relatively thick, impalpable red or maroon shales and clays.

This report presents the results of an investigation made on cross-bedding directions in the Triassic sandstone units cropping out in the area indicated in figure 1. The source areas for Upper Triassic sediments in this area are imperfectly known, and study of the abundant cross-beds appeared to offer the most reliable means of determining source areas by delineation of the regional direction of stream transport. Measurements of cross-bedding were initiated in Garza County. Supplementary data were obtained later on a more limited scale in Dickens, Borden, Scurry, Mitchell, and Howard counties.

The importance of cross-bedding studies in regional paleogeographic interpretation has been adequately evaluated as a result of investigations such as those made by Brett (1955), Katich (1953), McKee (1940), Olson and Potter (1954), Pelletier (1958), Pettijohn (1957), Potter and Olson (1954), Reiche (1938), Schwarzacher (1953), Shotten (1937), and Tanner (1955).

[1] Manuscript received November 18, 1959.

## STRATIGRAPHY

Upper Triassic sediments were deposited in a broad shallow syncline which is now largely covered by the Llano Estacado, a broad flat upland of about 20,000 square miles capped by indurated Tertiary sediments. Dockum units rest unconformably upon a dissected post-Permian terrane throughout the region and are unconformably overlain by Cretaceous and/or Tertiary rocks. The Triassic rocks range in thickness up to 450 ft in the areas of outcrop but thicken to 2100 ft or more toward a depocenter to the west in the subsurface beneath the Llano Estacado (Adkins, 1932).

The Dockum Group was first described and named by Cummins (1890) for exposures along Dockum Creek in Dickens County, Texas. Gould (1907) later divided the Dockum into a lower shale and an upper sandstone termed the Tecovas and Trujillo Formations, respectively. While these units can be recognized in Gould's area of study in the Canadian River valley and Palo Duro Canyon, the Triassic is not susceptible to such a simple subdivision south of Crosby County. Drake (1892) described sections in Howard and Mitchell counties made up of a lower sandy shale, central beds of sandstone, conglomerate and clay, and upper beds of sandy clay. For this same general area Hoots

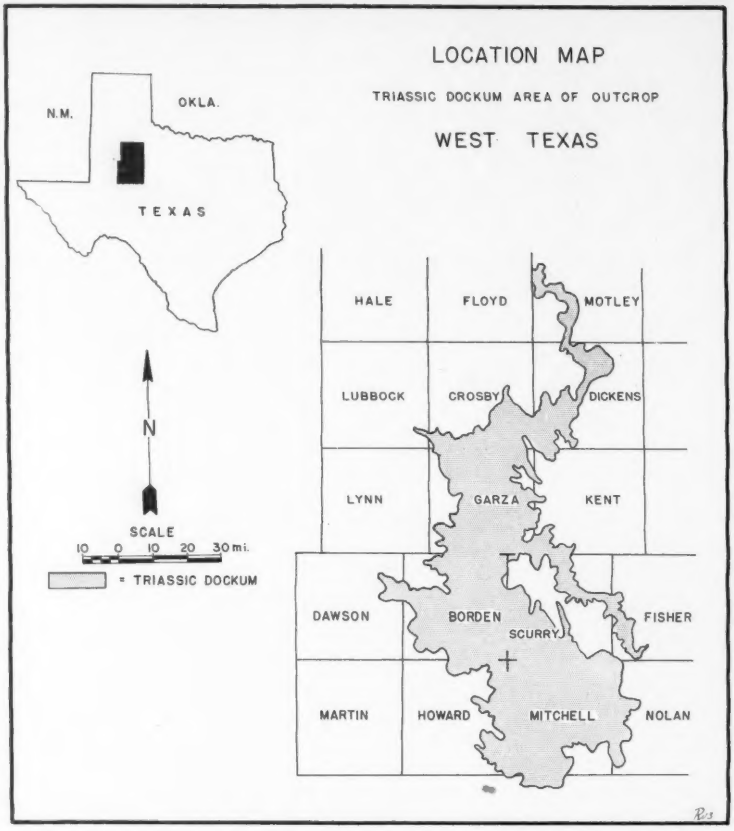


FIG. 1.—Location map of Upper Triassic outcrop.

(1925) found two formations, a lower red clay 275 ft thick with numerous sandstone units and an upper red clay 175 ft thick. Hoots claimed that his lower formation included Drake's basal clay and medial sandstones and that his upper clay is the equivalent to Drake's upper sandy clay. The difficulties encountered by these investigators are caused in part by rapid facies changes and an indeterminable amount of post-Triassic erosion. There are no criteria yet available that will establish the relationship between the multiple sandstone units in this southerly area with the Tecovas-Trujillo subdivision further to the north.

There is little published data regarding the stratigraphic sequence within the Dockum Group in the subsurface.

## SOURCE AREAS

Early investigators of the Triassic in west Texas visualized a flood plain type of environment with the sediments being transported by streams from highlands in south-central Colorado or the Wichita Mountain area in Oklahoma (Baker, 1915; Adkins, 1932). A mineralogical study by Sidwell (1945) appeared to substantiate the existence of a northwesterly source area, at least for that portion of the Triassic which crops

out in the Canadian River valley and adjacent portions of New Mexico. In contrast, Roth (1953) believed that Dockum conglomerates were derived from an area to the south and southwest of the Glass Mountains; his study was based upon study of the lithology of the pebbles and a fusilinid-bearing chert shingle in Motley County. Adams (1929) stated that the Triassic was deposited on a west and northwest sloping post-Permian surface because of the relative coarseness of the Triassic conglomerates along the eastern shelf area. Adams believed that the sediments were probably derived from sources to the east and southeast, although some contributions may have come from Colorado and northern New Mexico.

There can be little doubt that sediment contributions of varying volume emanated from several sources, considering the vast areal extent of the Triassic in this region. The data presented in this report, however, support the contention that major sediment contributions originated from the southeast and east of the Triassic depositional basin.

## PROCEDURE

Cross-bedded sandstones are sufficiently abundant, particularly in Garza County, that little difficulty was encountered in locating suitable outcrops. In most of the sandstones weathering has etched out the individual foreset beds so that the azimuth could be read directly with a Brunton pocket transit. The Triassic strata dip gently to the southeast but are essentially flat lying, and no correction for tilt was necessary.

Field work was initiated in Garza County. Fifty outcrops were visited at random, and a total of 192 measurements was recorded. A minimum of two measurements was made at each locality. Few outcrops exhibited more than one well-defined sedimentation unit although all sedimentation units were measured when present. The study was later extended by visiting 58 additional outcrops located in the six surrounding counties, raising the total of outcrops visited to 108. The grand total of measurements throughout the seven counties was 278, averaging 2.5 measurements per outcrop. The sandstone units studied represent the complete stratigraphic sequence of the Dockum Group from the basal sands above the Permo-Triassic unconformity to the highest sandstones immediately below the Triassic-Cretaceous (or Tertiary) boundary.

## ANALYSIS OF DATA

Figure 2 shows the location of individual outcrops with the average direction at each outcrop indicated by an arrow. In order to clarify the regional paleocurrent direction, a moving average grid was plotted (fig. 3). In constructing the moving average, the outcrops in all of four adjoining quadrants are summed vectorally, and the resultant vector is placed at the intersection of the four quadrants. By shifting the group of four one unit at a time east-west and north-south and repeating the procedure each outcrop is utilized four times. This method was applied by Pelletier (1958) in his study of paleocurrents in the Pocono Sandstone of Pennsylvania and Maryland.

The moving average thus constructed shows a preferred northwesterly direction for the majority of outcrops. This is particularly striking in Garza County. Further to the southeast the northwesterly direction still prevails but is less pronounced because of several reversals and a more westerly trend.

Rose diagrams in figure 4 indicate the range in fluctuation of the mean direction of current flow throughout Late Triassic time. Rose diagram (A) was constructed from measurements in the basal sands above the Permo-Triassic unconformity. Rose diagram (B) represents the uppermost sandstone units studied. Rose diagram (C) includes all measurements recorded throughout the stratigraphic section. The fluctuation of the mean between deposition of the lowermost units and uppermost units is only 7 degrees, indicating a stable tectonic environment. The duration of tectonic stability is perhaps more notable in the case of Lake Superior Precambrian quartzites studied by Pettijohn (1957). In that instance, cross-bedding studies indicated a stable paleocurrent system throughout the time necessary to deposit 20,000 feet of sediment. Pelletier found that between Upper Devonian and Lower Pennsylvanian sandstones the shift in mean cross-bed direction was 14 degrees.

The magnitude of the foreset dips was

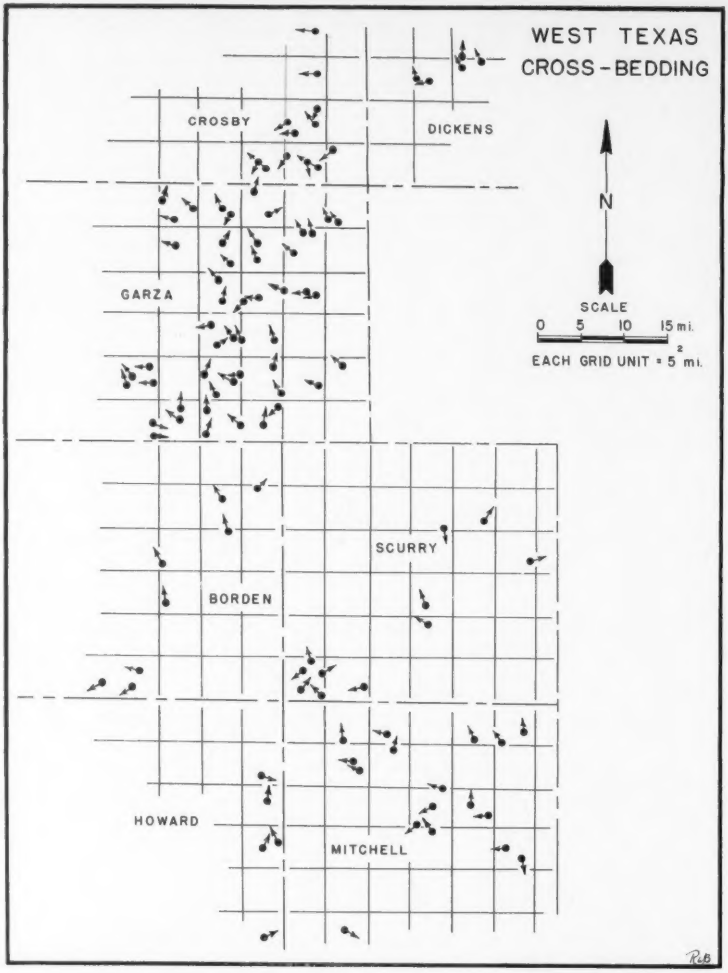


FIG. 2.—Average cross-bedding direction plotted at each outcrop locality.

measured, and the distribution of these is shown in the histogram of figure 5. The greatest number of dips fell within the 18 to 24 degree interval. These results are similar to those of Pettijohn (1957) and Potter and Olson (1954). The most common dips noted by Pelletier (1958) were within the 12 to 24 degree range. It appears that foreset dips of cross-beds characteristically lie within the 18–24 degree range.

## SUMMARY AND CONCLUSIONS

Cross-bedded sandstone units within the Upper Triassic Dockum Group of west Texas were studied in order to determine the mean direction of regional stream transport.

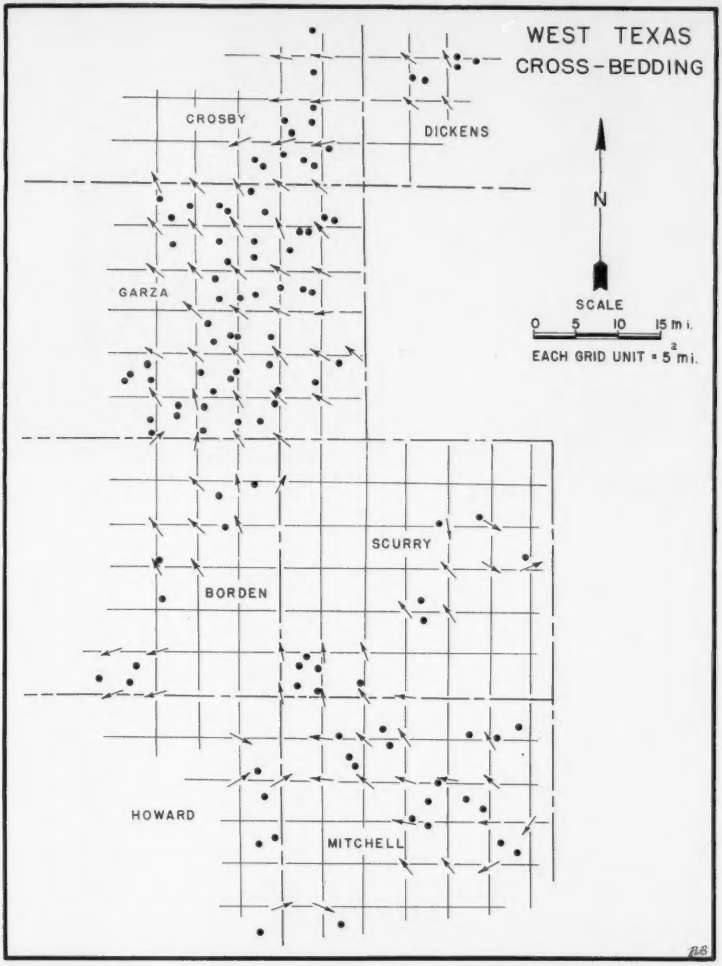


FIG. 3.—Moving average construction of Upper Triassic paleocurrent system.

Measurements were recorded at 108 outcrops scattered throughout 7 counties. By plotting the resulting dip directions on a moving average grid, a preferred direction of dip to the northwest is apparent. The inferred current flow was from the southeast to the northwest. The stability of the source area is indicated by the small fluctuation in mean cross-bedding direction between basal and upper sandstones within the Dockum Group. The results of this investigation lend support to Adam's (1929) contention that sources for the Triassic lay to the southeast and east. The writer suggests that the Llano Uplift region to the southeast in central Texas is a logical source area and was a principal contributor of sediments during Late Triassic time.

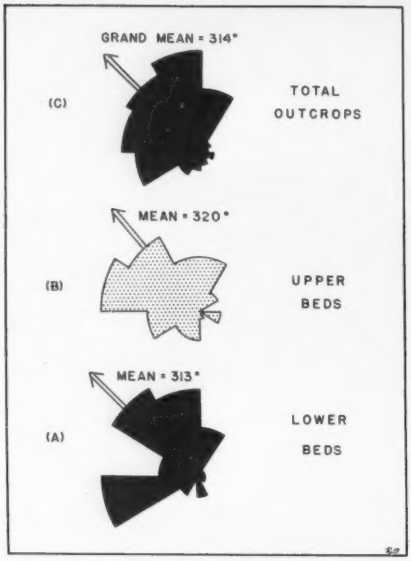


FIG. 4.—Rose diagram summary of Upper Triassic cross-bedding directions. Each arc is 30 degrees.

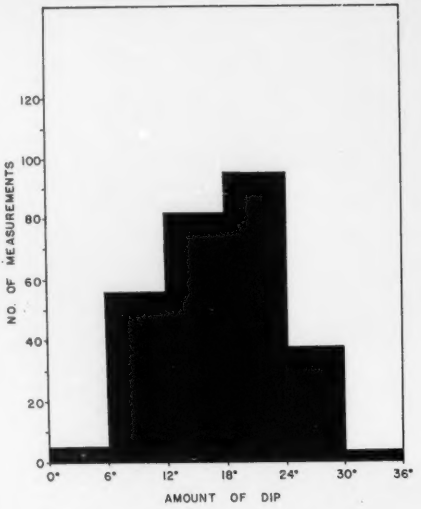


FIG. 5.—Histogram showing distribution of foreset dip magnitudes.

The more strongly preferred direction of cross-bedding dips in Garza County as compared to counties further to the southeast (and therefore closer to the inferred source) is at first anomalous. Pelletier's cross-bedding analysis (1958) showed opposite results, that is, increasing variation in a downstream direction reflecting the decline in stream gradients. It is possible that during Triassic sedimentation streams flowing over the eastern shelf area meandered frequently because of a relatively low gradient but further downstream were subjected to an increase in gradient at the hinge line of the eastern shelf of the Midland Basin.

ACKNOWLEDGMENTS

The writer wishes to thank Samuel W. Bishop and James C. Whitten of the Humble Oil and Refining Company who offered helpful comments during the course of the field work. Walter Wheeler of the University of North Carolina critically read the manuscript.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 466–470
SEPTEMBER, 1960

# THE RELATION BETWEEN SULFUR AND CARBON IN SEDIMENTS FROM THE ENGLISH LAKES[1]

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## ABSTRACT

Total sulfur and total carbon have been analyzed in fresh-water sediments from the English Lake District. The two elements are closely correlated, but carbon/sulfur ratios are lower in reducing than in oxidizing muds, and much lower in fertile than in infertile lakes. These ratios also vary considerably in sediment cores from a fertile and an infertile lake, being maximal during and just after the phase of most rapid increase in organic content.

## INTRODUCTION

Sulfur is a biophile element, and one whose behavior is strongly influenced by the oxidation-reduction potential of the environment (Mortimer, 1941, 1942). This paper treats the effect of three factors upon the relation between total sulfur and total carbon in fresh-water sediments: (1) oxidizing and reducing conditions, (2) lake fertility, and (3) time.

## THE LAKES

Five lakes were examined, as follows:

*Wastwater and Ennerdale Water.*—Unproductive deep lakes (79 and 45 m respectively), with waters poor in lime (0.09 m.equiv/1) and potash (0.01 m.equiv/1), and bottom waters in which oxygen is present throughout the year.

*Windermere North Basin.*—A deep lake (67 m) of moderate productivity, with water intermediate in lime (0.27 m.equiv/1) and low in potash (0.01 m.equiv/1), and bottom water which remains oxygenated at all times.

*Esthwaite Water.*—A shallow (16 m) productive lake, with water richer in lime (0.41 m.equiv/1) and potash (0.02 m. equiv/1), and bottom water which becomes completely de-oxygenated during the period of summer stratification.

*Priest Pot.*—A small shallow pond (4.5 m), very productive, and with the water richest in lime (0.62 m.equiv/1) and potash (0.08 m.equiv/1) among those studied, the bottom water becomes completely anaerobic during summer stratification.

Further data on these lakes and their sediments are given in an earlier paper (Gorham, 1960).

## METHODS

### Sampling

To study the influence of redox potential, samples were collected in October and November 1954, from Windermere and Esthwaite Water, at the surface and at about 5 and 10 cm beneath the surface of the mud. Since the lakes were isothermal at the time, the surface muds were distinctly oxidizing, as was evident from the presence of much reddish-brown ferric hydroxide in them. The samples from 5 and 10 cm depth were strongly reducing, and blackish in colour owing to the abundance of ferrous sulfide (cf Mortimer, 1941, 1942; Gorham, 1958).A series of six oxidate crusts, exceptionally rich in manganese and iron, was also collected at various times from the bottom of Windermere, together with two from Ullswater, a lake with somewhat similar characteristics.

The effect of lake fertility upon carbon/sulfur relations was examined by collecting samples of the top 5 cm of profundal mud from the five bodies of water mentioned above, under isothermal conditions during the spring of 1958.

The influence of time upon carbon and sulfur accumulation was studied in both a productive and an unproductive lake, Esthwaite Water and Ennerdale Water respectively. Sediment cores were taken

[1] Manuscript received November 6, 1959.

TABLE 1.—*Carbon and sulfur in oxidizing and reducing muds from Windermere and Esthwaite Water*

| Sample No.* | Lake | Redox Status† | Sample Depth, cm | Carbon | Sulfur | Carbon |
|---|---|---|---|---|---|---|
| | | | | % dry weight | | Sulfur |
| 1a | Windermere | O | surface | 14.7 | 0.39 | 37.7 |
| b | Windermere | R | 5 | 12.4 | 0.46 | 27.0 |
| c | Windermere | R | 10 | 12.2 | 0.47 | 26.0 |
| 2 | Windermere | O | surface | 8.3 | 0.26 | 31.9 |
| 3a | Windermere | O | surface | 8.7 | 0.23 | 37.8 |
| b | Windermere | R | 5 | 8.9 | 0.31 | 28.7 |
| c | Windermere | R | 10 | 8.1 | 0.29 | 27.9 |
| 4a | Esthwaite | O | surface | 11.5 | 0.33 | 34.8 |
| b | Esthwaite | R | 5 | 10.5 | 0.51 | 20.6 |
| c | Esthwaite | R | 10 | 8.4 | 0.41 | 20.5 |
| 5a | Esthwaite | O | surface | 11.6 | 0.44 | 26.4 |
| b | Esthwaite | R | 5 | 11.2 | 0.68 | 16.5 |
| c | Esthwaite | R | 10 | 9.9 | 0.47 | 21.1 |

* Samples 1 and 2 represent shallow water deposits, the others are profundal.
† O = oxidizing, R = reducing.

from the deepest parts of the lakes, by a new coring device (Mackereth, 1958; Smith, 1959) reaching into the glacial clays underlying the much more organic muds.

### Analysis

The samples were dried at 105–110° C. Total carbon was estimated gravimetrically by dry combustion, and since free carbonates are negligible in these muds the values represent organic material. Total sulfur was determined as sulfate by an ion-exchange technique (Mackereth, 1955), using a new oxidation method originated by F. J. H. Mackereth (in preparation).

## RESULTS

### The Influence of Redox Conditions

Table 1 provides data on oxidizing and reducing muds. In Windermere the surface sediment is continually oxidizing while in Esthwaite Water it becomes reduced each summer. Although the surface oxidizing muds are slightly richer in carbon, the subsurface reducing sediments are considerably richer in sulfur, so that the carbon/sulfur ratio averages 34 for the oxidizing muds and only 24 for the reducing samples. Such a result is to be expected, since sulfur tends to migrate from regions of high redox potential, where it may be oxidized and move as soluble sulfate, to regions of low potential where it is immobilized as insoluble ferrous sulfide.

The Windermere carbon/sulfur ratios are higher in both types of sediment, owing chiefly to lower sulfur contents. Furthermore, the difference in C/S ratio between oxidizing and reducing muds is greater in Esthwaite Water (avg. C/S = 31 for oxidizing and 20 for reducing samples) than in Windermere (C/S = 36 in oxidizing and 27 in reducing samples). The comparatively high sulfur level of the Esthwaite mud is not surprising, for it tends to be more strongly reducing than Windermere mud (Mortimer, 1942). This should favor the retention of sulfur, by inhibiting oxidative breakdown of organic sulfur and of ferrous sulfide to easily lost sulfate. Low redox potentials may also bring about considerable migration of sulfate-sulfur from the lake water into the mud, where it becomes reduced and precipitated as sulfide. The greater contrast in carbon/sulfur ratios of oxidizing and reducing sediments in Esthwaite Water agrees with the greater contrast of winter redox potential in these muds. It is, however, somewhat unexpected in view of the summer reduction of Esthwaite surface mud, and we must therefore presume that inorganic sulfur migrates fairly rapidly following the onset of oxidizing conditions there.

The most highly oxidized sediments found were the manganese-iron crusts collected in Windermere and Ullswater. These crusts, which appear similar to those investigated in considerable detail by Ljunggren (1953, 1955 a and b), were mostly

TABLE 2.—*Carbon, sulfur, iron and manganese in oxidate crusts from Windermere and Ullswater*

| Sample No. | Lake | Iron | Manganese | Carbon | Sulfur | Carbon |
|---|---|---|---|---|---|---|
| | | % dry weight | | | | Sulfur |
| 1* | Windermere | 38.6 | 7.0 | 0.8 | 0.030 | 27 |
| 2 | Windermere | 22.1 | 13.2 | 1.1 | 0.043 | 26 |
| 3 | Windermere | 16.2 | 13.2 | 1.3 | 0.048 | 27 |
| 4 | Windermere | 11.5 | 10.0 | 0.8 | 0.030 | 27 |
| 5 | Ullswater | 11.1 | 8.0 | 0.5 | 0.015 | 33 |
| 6 | Windermere | 8.5 | 2.3 | 0.9 | 0.028 | 32 |
| 7 | Ullswater (stony) | 5.9 | 2.9 | 0.4 | 0.010 | 40 |
| 8 | Windermere (clayey) | 7.2 | 1.4 | 0.3 | 0.005 | 60 |
| Avg. of 3 surface muds | Windermere | 6.13 | 1.14 | 10.57 | 0.293 | 36 |
| Avg. of 4 muds from 5 and 10 cm depth | Windermere | 4.58 | 0.22 | 10.40 | 0.383 | 27 |

* This sample consisted of very small fragments embedded in the mud where sample 2 was also collected. Samples 2–8 represent large crusts of several sq. cm area and roughly 0.5 to 3 or 4 cm thickness.

taken inadvertently by bottom dredges and nets, so that little can be said of their mode of occurrence, or whether they are present in others of the lakes studied. Analytical data for these samples are given in table 2, from which it is evident that they are very much poorer in both carbon and sulfur than are the lake muds, but considerably richer in manganese and iron. It is of interest that while carbon and sulfur values are higher in the crusts richer in iron and manganese, the ratios of these elements to total iron plus manganese exhibit a tendency to decline.

The positive correlation between sulfur and carbon in the oxidate crusts is extremely close. However, the ratio of carbon to sulfur varies greatly, being about 27 in the crusts richest in manganese, iron, carbon and sulfur, and reaching 60 in the somewhat clayey sample poorest in these elements. Rather strangely, the richest crusts have carbon/sulfur ratios similar to those of the reducing Windermere muds, and considerably lower than those of the oxidizing surface samples.

### The Influence of Lake Fertility

Both carbon and sulfur are more plentiful in sediments from productive lakes than in those from infertile situations, as shown in table 3. There is a very close positive correlation between the two elements, which may be due in part to their combination in organic matter. However, inorganic sulfides are certainly present, and free sulfur may also occur in these muds. It seems probable that fixation of these inorganic forms of sulfur will also be maximal in the most highly organic muds, since these tend to exhibit the greatest reducing intensity.

The infertile lakes have muds with high carbon/sulfur ratios (40 and 35 in Wast-

TABLE 3.—*Carbon and sulfur in profundal muds from lakes of differing fertility*

| Lake | Fertility Assessment | Carbon | Sulfur | Carbon |
|---|---|---|---|---|
| | | % dry weight of 0–5 cm sample | | Sulfur |
| Wastwater | Low | 7.2 | 0.18 | 40 |
| Ennerdale Water | Low | 6.6 | 0.19 | 35 |
| Windermere | Moderate | 8.7 | 0.37 | 24 |
| Esthwaite Water | High | 11.6 | 0.61 | 19 |
| Priest Pot | Very high | 18.9 | 1.22 | 15 |

TABLE 4.—*Carbon and sulfur in sediment cores from the deepest parts of Ennerdale and Esthwaite Waters*

| Ennerdale | | | | Esthwaite | | | |
|---|---|---|---|---|---|---|---|
| Depth, cm | Carbon | Sulfur | Carbon | Depth, cm | Carbon | Sulfur | Carbon |
| | % dry weight | | Sulfur | | % dry weight | | Sulfur |
| 16 | 5.9 | 0.24 | 25 | 20 | 8.0 | 0.42 | 19 |
| 43 | 5.4 | 0.19 | 28 | 50 | 6.6 | 0.35 | 19 |
| 70 | 11.6 | 0.31 | 36 | 80 | 8.4 | 0.40 | 21 |
| 124 | 7.6 | 0.27 | 28 | 110 | 7.1 | 0.41 | 17 |
| 178 | 8.2 | 0.29 | 28 | 140 | 7.2 | 0.42 | 17 |
| 200 | 7.1 | 0.28 | 25 | 170 | 8.6 | 0.39 | 22 |
| 232 | 8.6 | 0.30 | 29 | 200 | 7.7 | 0.41 | 19 |
| 250 | 9.4 | 0.32 | 29 | 250 | 9.1 | 0.58 | 16 |
| 286 | 9.4 | 0.29 | 32 | 280 | 8.7 | 0.54 | 16 |
| 300 | 8.5 | 0.27 | 31 | 310 | 8.6 | 0.56 | 15 |
| 350 | 8.4 | 0.24 | 35 | 340 | 9.5 | 0.59 | 16 |
| 400 | 7.4 | 0.18 | 41 | 380 | 9.5 | 0.32 | 30 |
| 450 | 3.9 | 0.11 | 35 | 420 | 9.3 | 0.40 | 23 |
| 500 | 5.5 | 0.15 | 36 | 460† | 4.5 | 0.59 | 8 |
| 550 | 2.5 | 0.075 | 33 | 570† | 2.2 | 0.17 | 13 |
| 600 | 2.2 | 0.085 | 26 | | | | |
| 650 | 1.5 | 0.053 | 29 | | | | |
| 690* | 0.7 | 0.025 | 26 | | | | |

* Just above glacial clay.
† Thin organic band in underlying glacial clay.

water and Ennerdale Water respectively), while the most productive lakes yield the lowest ratios (19 in Esthwaite Water and 15 in Priest Pot). This difference might be taken to reflect a lower sulfur content in the hard volcanic rocks surrounding the less fertile lakes as compared with the softer sedimentary rocks around the more fertile waters. However, it appears that while carbon/sulfur ratios of lake muds are much higher in the less fertile lakes, the ratios of dissolved bicarbonate-carbon to dissolved sulfate-sulfur in the lake waters are much lower in the less productive lakes, being only about 0.4 in Wastwater and Ennerdale Water as against 0.9 in Windermere, 1.2 in Esthwaite Water, and 2.6 in Priest Pot. It should be pointed out in this connection that only a small part of the total carbon and sulfur dissolved in the lake waters becomes fixed in the plankton (Lund, 1957), and still less must be sedimented to the bottom. Probably the main factor involved in bringing about low carbon/sulfur ratios in the muds of the productive lakes is their greater reducing intensity, which increases sulfur fixation as mentioned above.

### The Influence of Time

Some variation of carbon and sulfur accumulation in lake muds might be expected in the course of lake evolution, and is in fact exhibited by the sediments of both Ennerdale and Esthwaite Waters (table 4). In the unproductive Ennerdale site both carbon and sulfur increased relatively slowly in concentration, so that maximum values were not reached until over half the total depth of mud had accumulated; and there was also a distinct decline of both elements in the uppermost muds. However, the carbon/sulfur ratio varied only from 25 to 41, with the maximum values occurring during and just after the phase of most rapid increase in organic content.

In the productive Esthwaite Water a more complex picture is observed, and it must first be pointed out that the whole mud core exhibits exceptionally low percentages of organic matter. Most carbon values are below those recorded in tables 1 and 3, and other cores with almost double the carbon percentages have been collected from the depths of the same lake. However, the present core is of especial interest in

that while the carbon contents are similar to those of the Ennerdale core, the sulfur levels are markedly higher. The trend of carbon accumulation is also distinctly different, with the maximum percentages occurring only a little above the glacial clay. A subsequent decline in carbon is evident in Esthwaite as in Ennerdale Water. The carbon/sulfur ratios of the Esthwaite muds are particularly interesting, since they show marked fluctuations. Minimal values (8 and 13) are recorded for the thin organic bands found in the glacial clay, and a good deal of the sulfur in these samples is presumably derived from the pulverised sedimentary rock. The highest ratios (23 and 30) are observed in the first two mud samples above the glacial clay, immediately following the time of most rapid rise in organic content. Their ratios resemble the lower ones among the Ennerdale samples, and may perhaps be taken to suggest a phase of relatively oligotrophic conditions at this time, despite the high carbon levels. Unusually low proportions of chlorophyll derivatives in the organic matter of these two samples provide further evidence for this view (Gorham, unpublished, see also Gorham, 1960), and the exceptional accumulations of manganese at similar levels in other Esthwaite cores have been cited in support of the hypothesis that these earliest muds did not undergo reduction at the surface during summer (Mackereth, 1959).

## DISCUSSION

The main feature of the present results is the close relationship between total sulfur and total carbon in lake sediments. It is evident however that carbon/sulfur ratios may vary considerably in response to the three inter-related factors examined in this study—oxidation-reduction balance, fertility and time. Of particular interest is the fact that total sulfur exhibits a much wider range of variation between fertile and infertile lakes than does total carbon, the ratio of maximum to minimum concentration being about 7 in the former case and 3 in the latter. Even greater variation is however shown by the amounts of chlorophyll derivatives present in these muds (Gorham, 1960). From the scanty data reported, carbon/sulfur ratios appear to show promise as an aid to stratigraphic interpretation of sediment cores.

## ACKNOWLEDGEMENTS

This work was carried out while the writer was on the staff of the British Freshwater Biological Association, Ambleside, England. He is grateful to Mr. G. J. Thompson and the laboratory staff for collecting samples, and to Mr. F. J. H. Mackereth for advice and assistance with the carbon and sulfur analyses.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 471–480
FIGS. 1–7, SEPTEMBER, 1960

# SOME OBSERVATIONS OF RECENT OOLITES[1]

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## ABSTRACT

Oolite grains and oolitic sands are described from a newly recognized depositional area along the Texas Gulf Coast. These deposits occur in the high salinity lagoon environment of the Laguna Madre. They are associated with the wave exposed shorelines characterized by low rates of terrigenous sedimentation, high salinities, high summer temperatures, and active carbonate deposition. The descriptive details presented include comparisons with the oolite grains and oolitic sands from the Great Salt Lake and the Bahamas. Special reference is made to the radial, tangential, and unoriented crystallites within the oolitic layers from these separate localities. It is concluded that the Laguna Madre oolites are forming on the wave exposed shoreline by insolation, evaporation, and agitation of carbonate saturated waters. The internal arrangement of the individual aragonite crystallites within the oolitic layers seems dependent upon the rate of carbonate precipitation and the degree of mechanical reorientation by grain to grain rubbing.

For the past several years extensive studies have been made of the sediment distribution and faunal assemblages within the essentially brackish water bays of the central Texas coast (Shepard and Moore, 1955). These field and laboratory studies have been extended (Rusnak, 1960) to the hypersaline environment of the Laguna Madre along the south Texas coast (fig. 1).

The occurrence of oolites in the Laguna Madre sediments is one of the most significant characteristics of this hypersaline environment. There appears to be no other known area of present day oolite formation along the northern Gulf of Mexico Coast with the exception of the Florida coast and the adjoining Bahama Banks areas. For this reason, many samples were examined (fig. 2) in an attempt to evaluate the oolite distribution, their descriptive characteristics, and the environmental control. The earlier detailed and definitive studies by Eardley (1939) on the Great Salt Lake oolites and by Illing (1954) on the Bahaman oolites have given considerable insight into their formation (See also other recent studies by Carozzi, 1957; and Lucas, 1955).

Qualitative field observations indicated that the oolites and oolitic sands were associated with the wave-exposed mainland shoreline of the northern Laguna Madre and Baffin Bay and in particular at Point of Rocks. In these areas, the deposition of terrigenous material seemed to be at a minimum as was evidenced by the relatively thick accumulations of indigenous shell materials. The sediments are dominantly calcarenites in which the oolite constituent can amount to 50 percent or more of the total sample. A line of close-spaced samples was collected here perpendicular to the shoreline specifically for studies of oolite distribution relative to the shoreline. These were subsequently sized and examined for detailed counts of the constituents in each size grade. The quantitative results confirm the field observations and are shown in figure 3 with a general shore profile and the median diameter of each sample.[2]

The sample constituents were identified with the aid of a binocular microscope and were tabulated as terrigenous, oolites, superficial oolites, multiple oolites, shell fragments, calcareous aggregates, and composite grains consisting of oolites and terrigenous grains cemented to shell fragments (fig. 4). The terrigenous component consists of undifferentiated grains of quartz, feldspar,

[1] Contribution no. 273 from the Marine Laboratory, University of Miami. A large portion of this work was carried out at the Scripps Institution of Oceanography and was supported by a grant from the American Petroleum Institute, Project 51. The writer prefers to use the term *oolites* for the individual grains and the adjective *oolite* for the deposits (see Pettijohn, 1957, p. 95). Manuscript received November 20, 1959.

[2] All samples collected for this purpose were taken from the top 2 to 8 cm of the surface material and represent single lithic units.

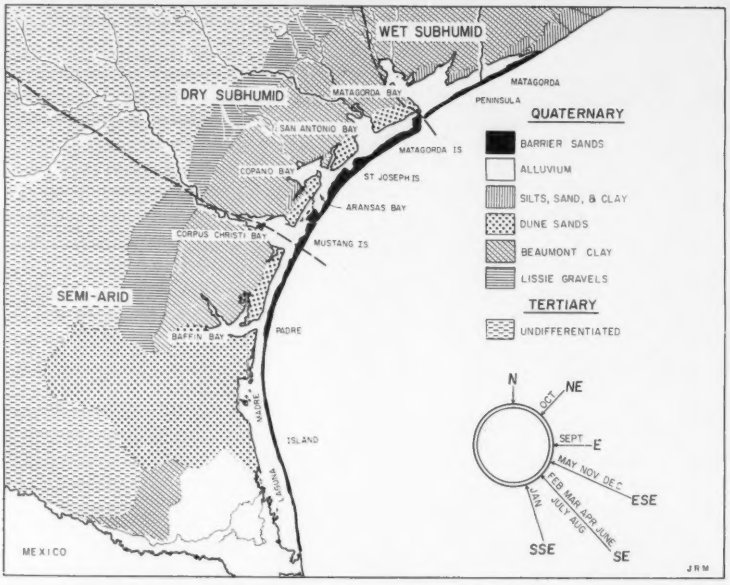


FIG. 1.—Geologic setting of the hypersaline Laguna Madre and its relation to the subhumid days to the north. Geology generalized from Duessen (1924) and Trowbridge (1923). Climatic boundaries after Thornthwaite (1948) and prevailing wind direction after Lohse (1955).

and chert, which do not show any carbonate coating whatsoever. The oolites consist of shiny, cream-colored, spherical to subspherical grains which appear to have relatively thick coatings of carbonate; superficial oolites show only a thin coating of shiny, cream-colored carbonate through which the nucleus protrudes or can be seen (fig. 5). Multiple oolites are made up of several grains which have been cemented together by an oolitic coating or layer of carbonate. Shell fragments consist mostly of broken mollusk fragments, commonly showing coatings of carbonate of varying thickness. Calcareous aggregates, ("lumps" of Illing, 1954) are generally irregularly shaped, shiny, cream-colored, and often possess attached sand and silt grains. The composite grains were identified as shell fragments with attached oolite and sand grains covered with a thin oolitic coating of carbonate cement. Table 1 demonstrates the distribution of these constituents in sample N at station 159 and is a representative sample of their occurrence in each size grade by $\frac{1}{2}\ \phi$ size increments of the coarse fraction.

A specific feature of all the carbonate-cemented or coated grains in the oolite zones is their shiny, cream-colored appearance as opposed to the white chalky appearance of many shells occurring in practically all parts of the lagoon except the oolite accumulation areas. Occasionally, a few shell fragments and composite grains display a white chalky surface in the oolite zones, but these are exceptions to the rule. Most commonly the white chalky material occurs in the protected cavities within the concave part of the shells and seldom on the convex surface.

The range of sizes represented by these deposits indicates that they are very well sorted to moderately well sorted ($\sigma\ \phi$ values ranging from 0.20 to 1.36 as derived by Inman's, 1952, graphic methods) and have median diameters of $\frac{1}{4}$ to $\frac{1}{2}$ mm. The oolites have size relations similar to those observed

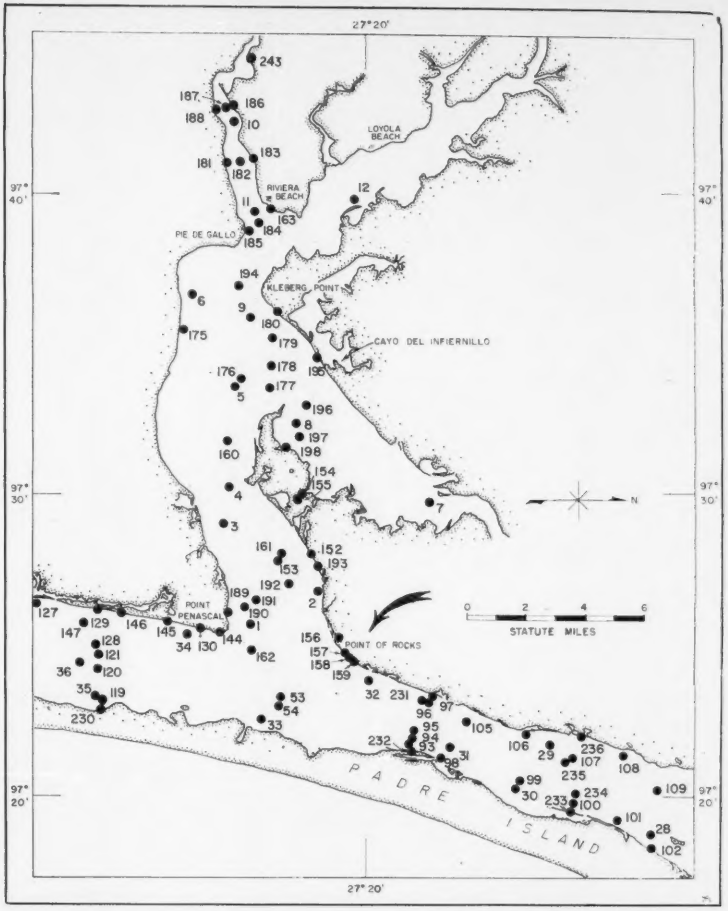


FIG. 2.—Station locations in Baffin Bay and the adjoining region of the northern Laguna Madre. Oolites occur along the mainland at "Point of Rocks." This shoreline is exposed to continuous wave action throughout most of the year because of the prevailing wind direction.

from the Great Salt Lake (Eardley, 1939; Carozzi, 1957), the Bahama Banks (Illing, 1954, p. 63, fig. 9), and the eastern Mediterranean (Lucas, 1955, p. 22); the oolites also have size relations similar to those universally noted for oolites throughout the geological column (Pettijohn, 1957, p. 95).

The greatest percentage of oolites occurs at the shoreline and the abundance peaks fall within the range of tides to be expected here. The abundance of oolites diminishes very rapidly in either direction away from the shoreline in all areas within the lagoon where oolites are found. This fact may indicate wave sorting of coarser materials on the beach or the actual point of oolite development. Several observations strongly suggest that the shoreline is indeed the area

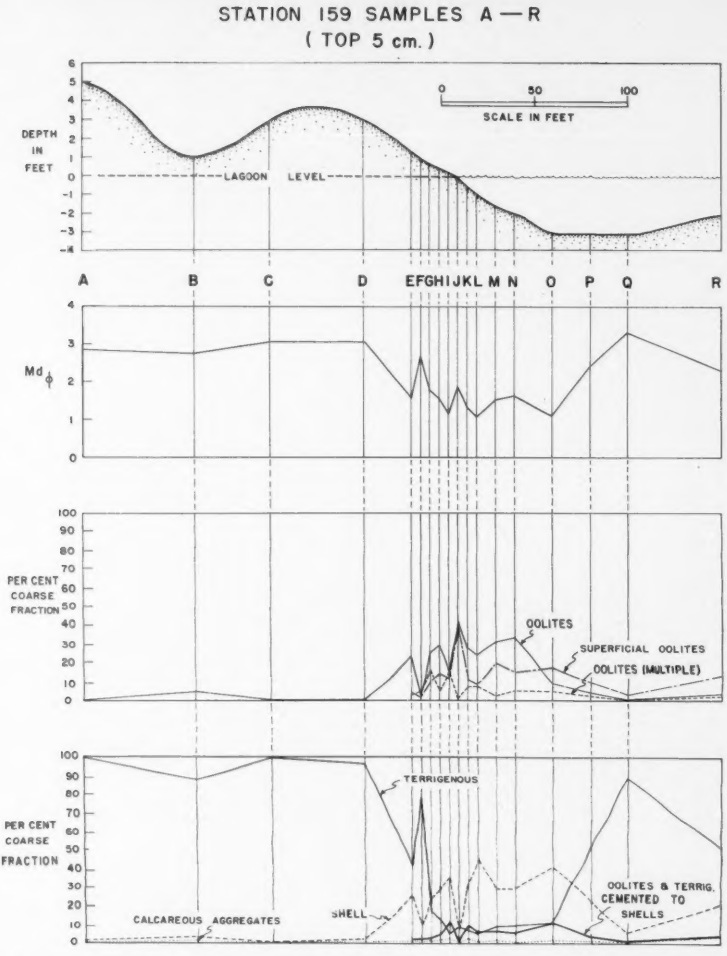


FIG. 3.—Quantitative results of constituent analyses obtained from samples collected perpendicular to the shoreline. Oolite abundance peaks occur at the shoreline and within the range of tides expected for this area. Tides are wind generated, for the most part.

of oolite growth as was found to be the case for the Great Salt Lake oolites by Eardley (1939).

There appears to be no outside source of oolites. Some oolites, apparently blown inland from the shoreline, are found in the small dunes bordering the lagoon, but where this has been observed the oolite grains appear to have lost their shiny surfaces by weathering and are whiter in color. Similarly, the dense Pleistocene beach rock occurring in this area exhibits very minor amounts of oolite grains in thin sections (Rusnak, 1960), but derivation of the recent oolites cannot be accounted for from this trace source. The restricted occurrence

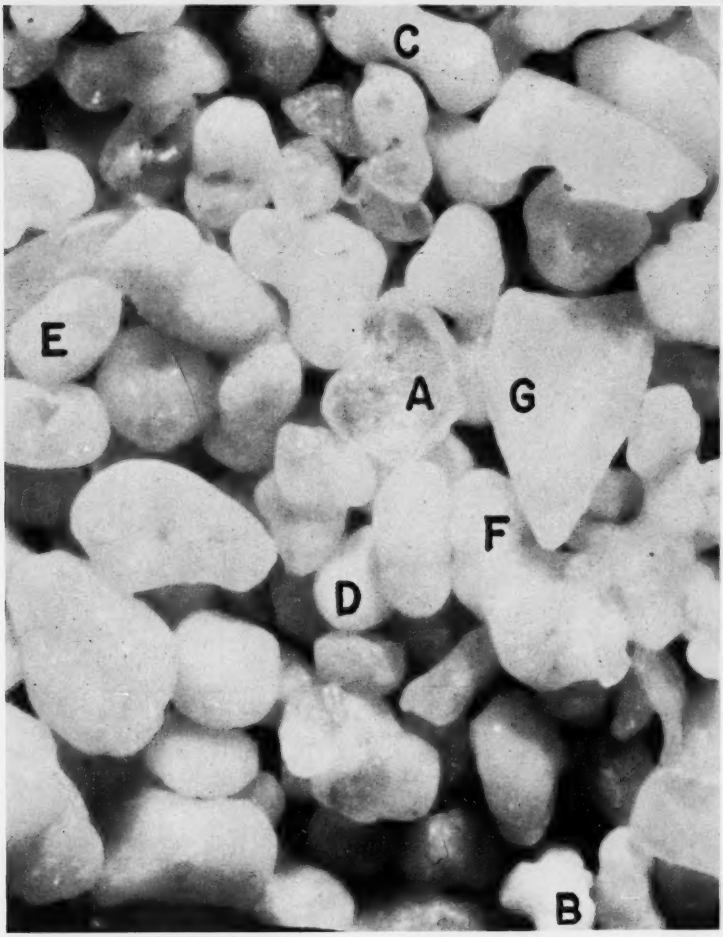


FIG. 4.—Photograph of the sample constituents occurring in the oolitic calcarenaceous deposits along the shoreline at "Point of Rocks." (A) Undifferentiated quartz, feldspar, chert, (B) Irregular carbonate grains with attached sand and silt particles, (C) Grains composed of several oolites cemented together by carbonate, (D) Single oolite grains with relatively thick carbonate layers, (E) Single oolite grains with thin carbonate layers through which the nucleus can be seen, (F) Shell fragments with attached oolites, sand, and silt grains covered by thin carbonate layers, and (G) Undifferentiated mollusk shell fragments.

of oolites suggests the shoreline as their place of formation. The strongest evidence for this contention, however, lies in the distribution of superficial oolites. If oolites were forming in any appreciably quantity throughout the lagoon, the concentration of these thinly coated, slightly smaller grains might be expected to remain reasonably constant throughout. Such is not the case. Rather, the superficial oolite distribution follows the abundance distribution pattern exhibited by the oolites, being most abun-

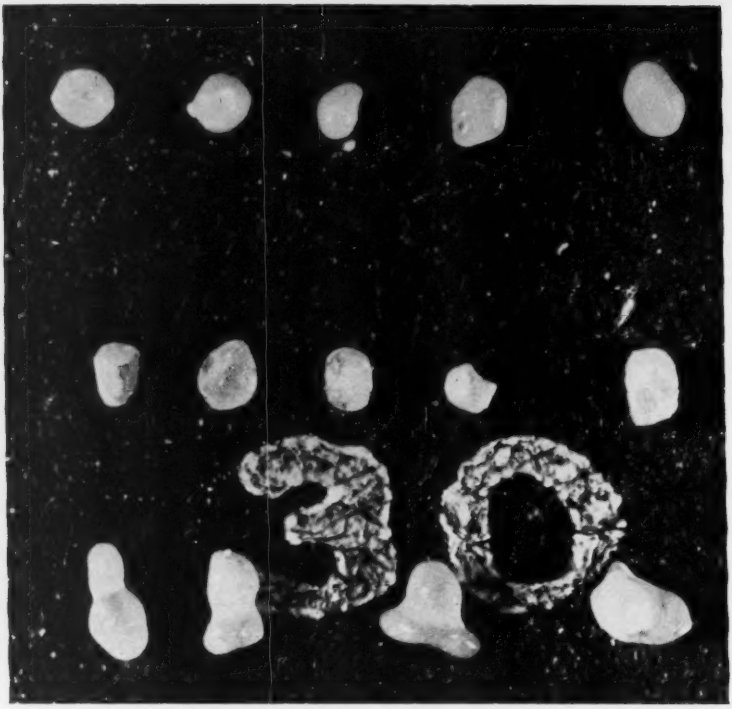

FIG. 5.—Photograph of oolite grains. Upper row consists of well developed single oolite grains. Middle row consists of superficial oolite grains. Lower row consists of multiple oolites made up of two or more oolites cemented together by aragonite oolitic layers.

dant in the highly agitated zone of the shore. In this zone, considerable amounts of precipitated carbonate are in evidence, for example, the abundant presence of carbonate coated shell fragments and the formation of beach rock (Rusnak, 1960).[3] Hence, the formation of oolites agrees with the "classical theory" of oolite origin which states that oolites grow in strongly agitated waters saturated with calcium carbonate (Eardley, 1939; Illing, 1954).

The Salt Lake oolites consist of a detrital mineral grain or fecal pellet nucleus surrounded by a series of concentric envelopes of aragonite and calcite with occluded clay (Eardley, 1939). In thin-section the concentric bands have mainly a *radial arrangement* of acicular carbonate crystallites which exhibit a pseudo-interference figure of the uniaxial type, expressed as a black cross under crossed micols. The Bahaman oolites, on the other hand, consist of a carbonate grain (carbonate aggregate or shell fragment) nucleus surrounded by a series of concentric aragonite envelopes in which the constituent acicular crystallites are arranged *tangentially* (Illing, 1954). These also exhibit a uniaxial pseudo-interference figure. In both of these oolite examples, some of the aragonite layers exhibit an *unoriented* aggregate structure of the acicular crystallites. Thus, several forms of

[3] Insolation and evaporation are believed to be the primary causes of aragonite precipitation, although $CO_2$ uptake through biologic control (here mostly by blue-green filamentous algae, Rusnak, 1960) obviously helps to promote precipitation.

TABLE 1.—*Coarse fraction constituent composition LM-159N*

| Phi Size Grade | Weight percent | Terrigenous a | Calcareous Aggregate b | Multiple Oolites c | Oolites d | Superficial Oolites e | Composite grains f | Mollusk Shells g | Millimeter size grade |
|---|---|---|---|---|---|---|---|---|---|
| <0 | 7.6 | | 2 | | | | 13 | 85 | >1.00 |
| 0– .5 | 7.7 | | | 2 | | | 25 | 73 | 1.00 –0.707 |
| .5–1.0 | 10.4 | | | 18 | 3 | 1 | 14 | 64 | 0.707–0.500 |
| 1.0–1.5 | 19.6 | 6 | | 12 | 45 | 4 | 3 | 30 | 0.500–0.354 |
| 1.5–2.0 | 40.7 | 7 | | 1 | 60 | 30 | Tr. | 2 | 0.354–0.250 |
| 2.0–2.5 | 9.2 | 11 | | Tr. | 18 | 71 | | Tr. | 0.250–0.177 |
| 2.5–3.0 | 1.8 | 81 | | | 2 | 17 | | 1 | 0.177–0.125 |
| 3.0–3.5 | 1.0 | 85 | | | 1 | 13 | | 1 | 0.125–0.088 |
| 3.5–4.0 | 0.1 | 88 | | | 1 | 10 | | 1 | 0.088–0.062 |
| >4.0 | 1.7 | | | | | | | | <0.062 |
| Percent Coarse Fraction | 100.0 | 9.8 | .2 | 5.5 | 5.7 | 28.9 | 15.7 | 34.2 | |
| Percent Whole Sample | 98.3 | 9.7 | .2 | 5.4 | 5.6 | 28.5 | 15.4 | 33.5 | |

a. Undifferentiated quartz, feldspar, chert.
b. Irregular carbonate grains with attached sand and silt particles.
c. Grains composed of several oolites cemented together by carbonate.
d. Single oolite grains with relatively thick carbonate layers.
e. Single oolite grains with thin carbonate layers through which the nucleus could be seen.
f. Shell fragments with attached oolites, sand, and silt grains covered by thin carbonate layers.
g. Undifferentiated mollusk shell fragments.

crystallite orientation are possible within oolites.[4] At the present time, there seems to be no satisfactory experimental or theoretical explanation for the control of the crystallite orientation during the primary deposition of the oolite layers. Sorby (1879, p. 74), however, believed the tangentially oriented layers were caused by the mechanical accumulation of crystallites on rolling grains in a manner similar to the build up of a snowball. Recrystallization and inversion of aragonite to calcite has been used to explain the radial structures which intersect the layer boundaries observed in older oolite (for example, Eardley, 1939, and Illing, 1954).

The internal structure of the Laguna Madre oolites, as observed in thin section, consists of a detrital nucleus surrounded by a series of up to 8 or 10 concentric aragonite (confirmed by X-ray analyses) layers (fig. 6). This structure was compared with thin sections of Great Salt Lake and Bahaman oolites. The separate carbonate layers of the Laguna Madre oolites include those with crystallites which are oriented radially, those which are oriented tangentially, and those which are unoriented (fig. 7). The unoriented layers are the most common and contain abundant occluded clay. Both the tangentially and radially oriented layers exhibit a pseudo-interference figure, but in the radial layers it is less distinct. In the unoriented layers, the figure is absent. Optically, the constituent crystals have a negative elongation with indices of refraction consistent with aragonite. These minute crystals occur mostly in the size range of one to five microns. The individual concentric envelopes show no systematic outward progression of oriented to unoriented crystallites from the nucleus to the outermost shell. The layers can alternate from unoriented to radially oriented or tangentially oriented envelopes in any ar-

[4] Only the original or primary structural features are considered here. The alteration of these primary structures by diagenesis has been discussed, for example, by Eardley (1939) and Illing (1954).

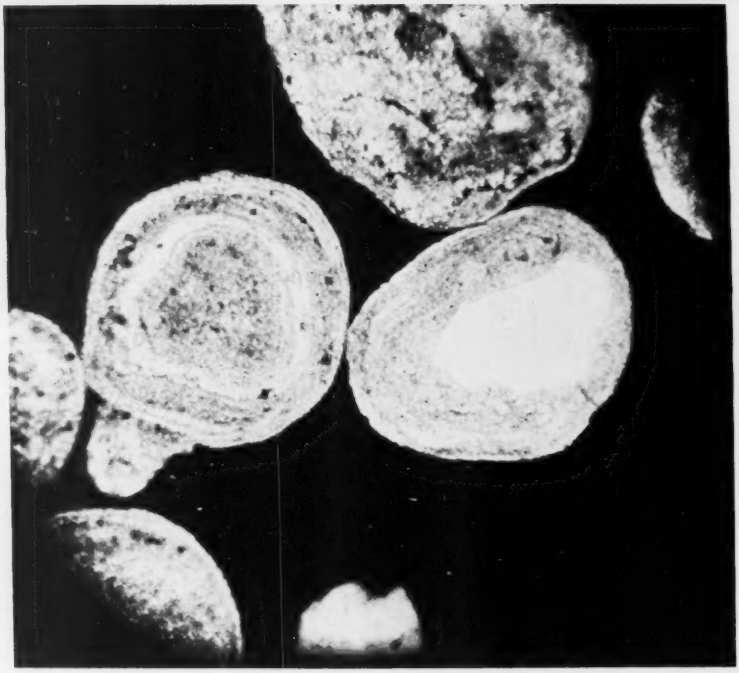

FIG. 6.—Laguna Madre oolites in thin section (polarized light). The aragonite layers surround both the grain with a detrital quartz nucleus and the grain with a carbonate aggregate nucleus, in this example.

rangement. Each layer appears to possess only one kind of internal orientation, and the "growth rings" appear not to be intersected in any manner which might indicate recrystallization. This observation suggests that the degree or kind of orientation is dependent upon the primary chemical factors and degree of agitation within the precipitating medium rather than diagenesis. If it were a diagenetic phenomenon, the older more internal layers might be expected to show the greatest alteration and the youngest layers to show the least. A systematic progression from the nucleus outward might be expected, but none was observed.

The rate of carbonate precipitation and mechanical reorientation may very well be the controlling factors of primary crystallite orientation within oolites. With very rapid precipitation, individual needles may not have an opportunity to become preferentially oriented on the nucleus and result in unoriented carbonate deposition. With a less rapid rate of precipitation they may become oriented radially as is demonstrated in artificially precipitated oolites or spherulites (see for example Monaghan and Lytle, 1956; LaLou, 1957). Where the precipitation rate is very slow, the crystallites could become attached tangentially to the nucleus by rolling or agitation, as originally suggested by Sorby (1879), or they could become bent by mechanical rubbing from an initial radial attachment to a tangential growth position. Thus, the degree of saturation and mechanical agitation of the precipitating medium could be responsible for the orientation of crystallites in oolites. It is significant that the common orientation

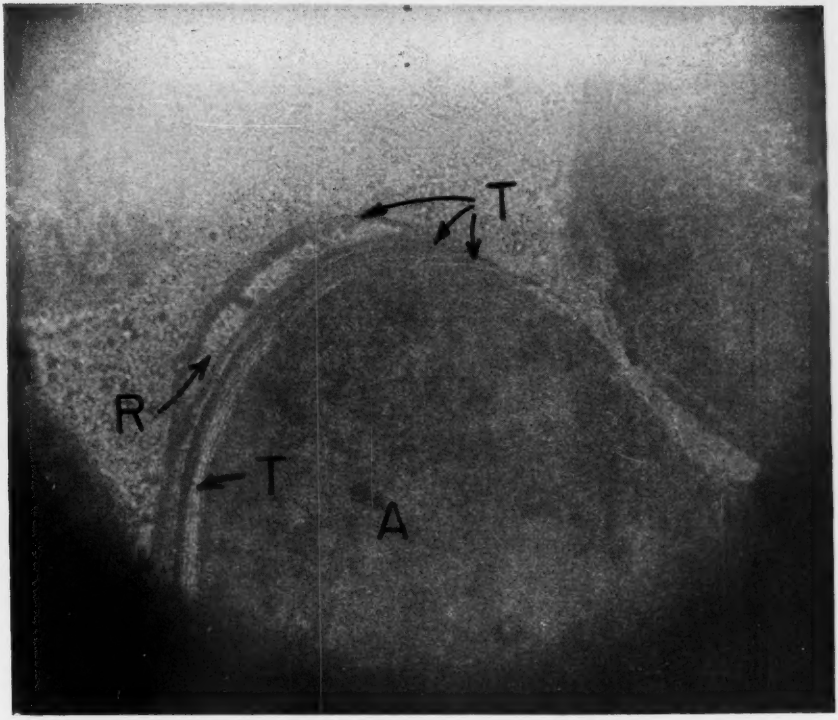


FIG. 7.—Oolite structure under high magnification (plane polarized light). This Laguna Madre oolite exhibits (T) tangential, (R) radial, and (A) aggregate crystallite orientations within the concentric aragonite envelopes and in the nucleus.

of growing aragonite crystallites in beachrock is radial to the nuclei (see, for example, Ginsburg, 1953; Rusnak, 1960). In beachrock there is little or no mechanical orientation since the shells are not in motion and the cement fills voids in the deposit (see also the void filling of "lumps" described by Illing, 1954). Proceeding with this argument, one readily concludes that tangentially oriented oolite layers must be subjected to relatively high crystallite strain during the bending process. These strained crystallites may thus be more susceptible to recrystallization by diagenetic processes in response to a release of acquired strain.

The Laguna Madre waters seem to have an intermediate salinity between that of the Great Salt Lake and the Bahama Banks waters, on the average. Because calcium carbonate solubility is partly dependent upon salinity (Revelle and Fairbridge, 1957), one would expect, with other controls being equal, that the Laguna Madre oolites should have internal structural and mineralogical characteristics common to both of the other environments. Thus, as saturation conditions fluctuate, there need not be a systematic progression of oriented to unoriented layers. The phenomenon can depend entirely upon the amount of change in the chemical environment (either diurnally or seasonally) and the degree of agitation during primary deposition of the oolites.

It is concluded that the Laguna Madre oolites are forming on the shoreline by

insolation, evaporation, and agitation of carbonate saturated waters. Moreover, the internal arrangement of the individual carbonate crystallites seems dependent upon the rate of carbonate precipitation; (1) with very rapid precipitation the carbonate is dumped in mass around a nucleus, (2) with slower precipitation and weak agitation the crystallites orient themselves radially, and (3) with very slow precipitation the crystallites become mechanically oriented by strong agitation and grain to grain rubbing from an initial radial attachment to a tangential growth position.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 481–485
FIGS. 1–4, SEPTEMBER, 1960

# SHALLOW WATER RIPPLE MARK VARIETIES[1]

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## ABSTRACT

Eight distinctive but little-known ripple mark varieties have been recognized in shallow water (less than one meter deep) along the Florida panhandle coast. Three additional well-known varieties have been seen in water up to 20 meters deep. The shallow-water varieties are thought to be caused by falling water level, direct wind effects, or a combination of two or more wave sets. One of these varieties, a symmetrical ripple mark, is shown to deviate in plan by as much as nearly 45° from the waves which produce it. Since this latter type is difficult to identify, its lithified counterparts might be easily misinterpreted.

During the past five years the present writer and his students have carried on a research program in near-shore sedimentology and morphology, largely under the sponsorship of the Research Council of Florida State University. This program has been concerned chiefly with sediments and landforms in a strip approximately 30 km wide, having the Florida panhandle shoreline as its approximate median (fig. 1). Part of the work has been reported by Brenneman and Tanner (1958), Waskom (1958), Vause (1959), and Tanner (1959). Other portions, particularly those dealing with the microfauna, are as yet unpublished.

In the course of this long-range study, many observations of ripple marks have been made. The variety of ripple marks which have been observed is now sufficiently great to warrant a preliminary description.

All of the ripple marks described here were found in water having a depth between a few millimeters and several tens of meters. Shallow-water ripple marks were studied on foot; deeper water ripple marks were studied by SCUBA-equipped divers operating from a boat. The best examples of the most interesting shallow-water varieties have been seen near Mashe's Island and Bald Point (fig. 1).

The ripple marks observed to date can be classified as follows:

A. Elementary ripple marks.
   1. Simple ripple marks.
      a. Wave-formed.
      b. Current-formed.
   2. Compound ripple marks.
      a. Regular.
      b. Irregular.

B. Flat-topped ripple marks.
   1. Simple.
   2. Compound.

C. Curved ripple marks.

D. Very short ripple marks.

E. Windrow ridges.

F. Wash-over crescents.

G. Composite ripple marks.

The elementary ripple marks are those generally described by many workers. They include the wave-formed

$$\left(\text{symmetric; Ripple Symmetry Index} = \frac{\text{crest-to-trough}}{\text{trough-to-crest}} = 1\right)$$

and the current-formed (asymmetric; RSI > 1) varieties. They may appear in simple sets or in compound sets (that is, two or more simple sets crossing at some definite angle). They also may appear as "tad-pole nests," small irregular ripple marks making an erratic polygonal pattern. The latter variety has been observed only in water a few centimeters deep and the more regular compound sets in water up to 20 m deep. Description and discussion of most of the common combinations may be found in several references (for example, Evans, 1941, 1942, 1943, 1949; Shrock, 1948).

Flat-topped ripple marks have been described by Wegner (1932), Shrock (1948), McKee (1957), and Tanner (1958). Those

[1] Manuscript received November 2, 1959.

FIG. 1.—Map of Florida coast south and southwest of Tallahassee, showing Bald Point and Mashe's Island.

reported herein were found exclusively in water only 2 or 3 cm deep; they had obviously been planed off during a fall in water level (that is, falling tide). The simple variety is sketched in figure 2-A, and the compound variety in figure 2-B. Lithified examples of simple flat-topped ripple marks have been seen in the (Pennsylvanian) Wewoka Formation of central Oklahoma, the (Pennsylvanian) Jackfork Formation of southeastern Oklahoma, and the (Jurassic) Morrison Formation of western Colorado. Lithified examples of compound flat-topped ripple marks have been observed in the (Pennsylvanian) lower Satanka sequence of north central Colorado (in the vicinity of Horsetooth Reservoir), and in the (Jurassic) Morrison Formation of western Colorado.

Wave-length, amplitude, ripple index, and ripple symmetry index data are apparently useless for flat-topped ripple marks.

Curved ripple marks occur commonly where two or more sets of simple ripple marks cross or intersect, and may be either systematic (fig. 2-G) or non-systematic (fig. 2-F). They should not be confused with the irregular pattern colloquially described as "tad-pole nests." Curved ripple marks may be related to windrow ridges (see below), may resemble miniature barchans, or may develop in obvious interference patterns (fig. 2-F). They have been observed, to date, in water generally less than 20 cm deep and at no place more than 1 m deep.

The category, very short ripple marks, refers to simple forms having wave-lengths of approximately 2 or 3 cm. They do not differ, other than in size, from the elementary ripple marks listed above. This size difference, however, is thought to be related directly to water depth; of the observed very short ripple marks none occurred in water more than a few tens of centimeters deep (Kindle, 1917). Lithified examples have been studied in the (Pennsylvanian) lower Satanka of north central Colorado, the (Permian) Yeso Formation of the southern Sangre de Cristo mountains of north central New Mexico, the (Jurassic) Morrison Formation of western Colorado, and the (Jurassic) Morrison Formation of north central Colorado.

Windrow ridges were studied in water less than 1 m deep. In each instance the windrow ridges, looking more or less like simple ripple marks, lay parallel with and directly beneath windrows on the water surface and cut angularly across a ripple mark field of some other kind. The most obvious windrow ridges were widely-spaced and crossed simple ripple mark fields at angles of 70° to 110° (fig. 2-E). Each ridge was tapered, having the narrowest portion downwind from the widest and highest portion and a slight step where the downwind end merged into the next ripple mark. From such ridges it would be possible to infer the wind direction.

Other windrow ridges occur in combination with curved ripple marks (fig. 2-G). In the example sketched the straight ripples are the windrow ridges, and the wind is blowing from left to right. The curvature of the other ripples in the same sketch is due to the presence of two wave sets, one advancing from the upper left hand corner of the sketch and the other from the lower left hand corner. These curved ripples are similar to those shown in figure 2–F.

Windrow ridges crossing curved ripple fields were generally a few tens of centimeters apart. The ridges might stand either higher or lower than the associated ripple marks.

In ground plan wash-over crescents look

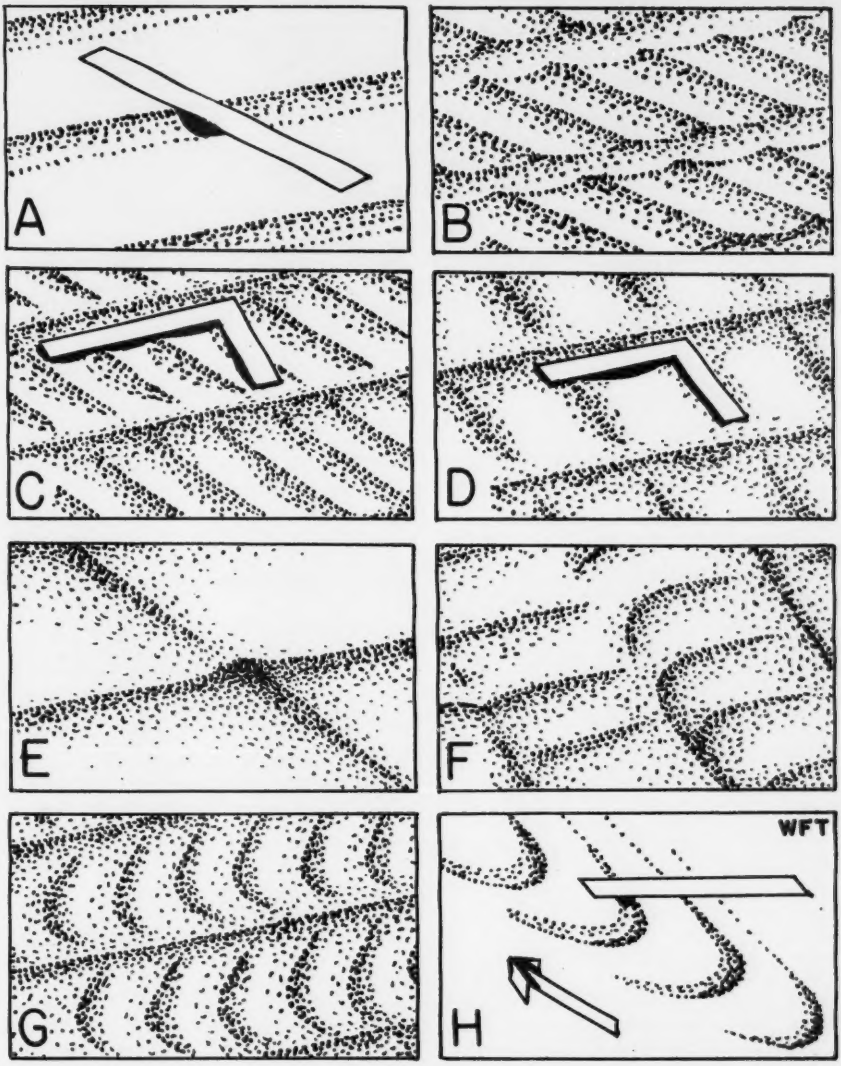


FIG. 2.—Sketch of eight ripple mark patterns actually observed. A, flat-topped ripple marks; note shadow cast by ruler. B, wave ripple marks formed on flat tops. C, compound ripple marks (both wave-formed) of two different wave lengths, not at right angles. D, compound ripple marks (both wave-formed) at right angles to each other. E, windrow ridges crossing simple ripple marks; wind (and windrows) move from upper left to lower right. F, curved ripple marks apparently formed in part by vector addition of particle orbits of two different wave sets. G, curved ripple marks caused by two different wave sets, in between long, straight windrow ridges. H, crescentic depressions caused by washover effects; water moves with the arrow.

like tiny barchans (fig. 2-H) but actually are coalesced flat-topped ripple marks having crescentic depressions between them. They have been observed on sand bars which were awash during falling tide. The troughs are generally approximately 1 cm deep and the crescents are a few tens of cm from tip to tip and some tens of cm apart. As is true of wind-formed barchans, the horns of the crescent point down-current.

The wash-over crescents grade up-current first into flat-topped ripple marks and farther up-current into simple ripple marks. Down-current they pass into smooth sand surfaces which are barely awash. In some instances the points may extend for a meter or more, giving rise to a diamond pattern on the sand. An additional pattern of rather erratic nature may develop from breaker-foam, deposited in lines which are roughly parallel with the water's edge.

Composite ripple marks are those formed by two intersecting sets of waves, neither of which is parallel with the ripple mark crests. In this respect they differ from compound ripple marks where two intersecting sets of waves produce two intersecting sets of ripple marks. Composite ripple marks occur in single sets only, with all ripple marks essentially parallel with each other and not parallel with the waves. In general appearance they resemble ordinary simple ripple marks; it occurs to the writer that they probably have been seen in the lithified rocks and interpreted as the common variety with attendant erroneous inferences as to wave patterns. They do have distinctive characteristics, however, which may permit positive identification.

Composite ripple marks have not been sketched in figure 2, but a low-angle photograph (with considerable distortion) appears as figure 3, and a diagram is given as figure 4. The ripple field in the photograph is covered by water approximately 20 cm deep. The low angle and the distortion caused by the uneven surface of the water make the ripple marks seem less perfectly parallel than they really are. The ripple mark crests form angles of approximately 40° with the wave crests. This situation is diagrammed in figure 4. Where ripples of this type occur within an obtuse angle between wave sets, the ripple marks are oriented at 90° to those in the sketch; in other words, the ripple mark crests bisect, roughly, the acute angle between waves. This is apparently a matter of vector addition of water particle orbits, as shown in the diagram.

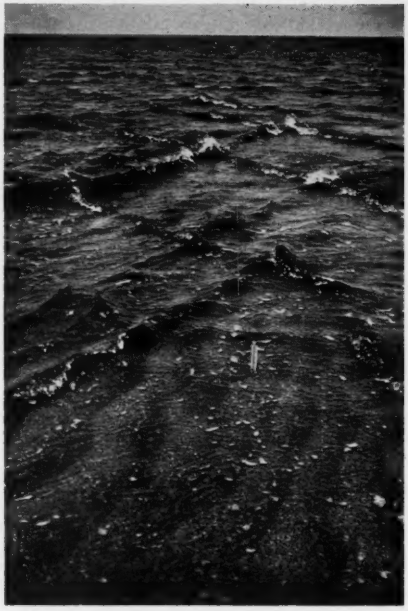

FIG. 3.—Composite ripple marks, apparently caused by vector addition of particle orbits of two different wave sets. The plastic ruler near the center of the picture indicates a water depth of approximately 20 cm. The ripple marks are actually straight and parallel, but the low camera angle and the uneven water surface combine to make them look irregular. The ripple marks bisect the acute angle between the two wave sets.

Wave lengths for composite ripple marks are commonly between 5 and 10 cm; amplitudes are generally 10 mm or less; and ripple symmetry index is one. The quartz sand of the area has median diameters between 2.8 and 1.5, and percentile deviations between 0.40 and 1.45. There is a minor admixture of fine shell hash.

Composite ripple marks are best developed on shoals where wave refraction produces two wave sets not quite 90° from each other. The case where the angle is precisely 90° has not been studied in operation yet, but it is reasoned that a much less regular

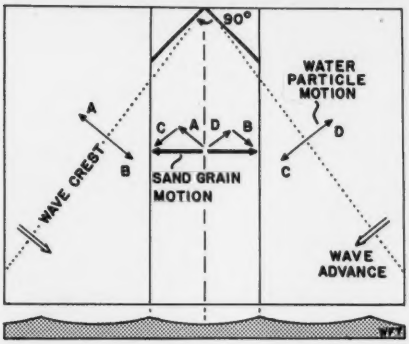


FIG. 4.—Proposed vector addition, to account for the fact that composite ripple marks deviate from water wave crests by approximately 40°. All four motions (A, B, C, D) are equally likely on the sand surface where wave orbits are flattened to simple oscillations, but the most efficient combination of motions yields $A+C$ and $B+D$; hence the ripple marks bisect the acute angle between the wave crests. The combinations $A+D$ and $B+C$ would be less efficient and therefore do not appear. Where the angle between wave crests is very close to 45°, the resulting pattern is neither so simple nor so easily defined.

ripple mark pattern would appear. Where the angle between wave crests departs from 90°, a field of apparently simple ripple marks occurs. On both sides of the shoal, as the ripple mark field passes into deeper water, the composite ripple marks grade into genuine simple ripple marks which actually parallel the wave crests. The simple ripple marks on both sides of the shoal under ideal conditions form two fields which, when taken together, yield a chevron pattern having the open end toward the shore. If all three components can be observed (the two simple ripple mark fields and the composite ripple marks in between), it may be possible to identify the latter.

A second identifying characteristic is the local offset, whereby each ripple mark crest is displaced one ripple mark wave length either to the left or right. These offsets occur at intervals of about 5 to 20 times the ripple mark wave length. No bifurcation is associated with offsets of this type, but rather several ripple marks are offset "in unison." A rather broad exposure would be necessary, however, for the field geologist to find and properly interpret these very regular but widely-spaced features.

Eleven varieties of ripple marks have been described. With the exception of the two types of simple ripple marks and the regular compound ripple marks all have been observed exclusively in shallow water (depths of 1 m or less). With the exception of the composite ripple marks, all can be recognized fairly easily. The latter, however, may lead to highly erroneous paleogeographic interpretations.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 486–489
FIGS. 1–2, SEPTEMBER, 1960

# COMPUTED SEDIMENT GRAIN SURFACE AREAS[1]

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## ABSTRACT

Approximate sediment grain surface areas were computed for 109 natural sediment samples ranging from sands to clays. This was done by assuming all particles to be spherical and summing the areas for each size fraction in the grain size analysis, where the average diameter for the size fraction was used for all grains in the fraction. The data are presented both as area per unit mass of dry sediment and as area per unit volume of wet sediment. Comparison of the results with theoretical curves for uniform spheres indicates the useful range of these computations.

Data on total sediment grain surface areas can be useful in certain geochemical and geophysical applications. In the field of acoustics, for example, it is of interest because the absorption of elastic wave energy in a mixture of grains and fluid is mainly the result of viscous heat losses occurring along grain surfaces. In a study of the absorption of sound in sediments in relation to other sediment physical properties (Shumway, 1960), surface areas were computed from sediment grain size fractions, assuming spherical particles, for more than 100 sediment samples ranging from sands to clays. These data are presented here (figs. 1, 2) to show the general range of values to be expected and the relation of surface area to median grain diameter.

The total surface area for an aggregate of particles of uniform size and shape varies inversely with linear dimension of the particles if the total particle volume remains constant. Natural sediments are composed of particles of various sizes and shapes, hence close agreement with this inverse relationship is not to be expected. However, because the grain size distributions of natural sediments usually have pronounced central tendencies, generally it is true that the smaller the median diameter (or other measure of central tendency), the larger the surface area.

A variety of methods have been developed for determining surface areas of finely divided material (Cadle, 1955; Orr and Dalla Valle, 1959). The different methods serve various purposes; for example, surface areas measured by chemical adsorption techniques are more useful for chemical considerations than they are for situations involving fluid flow where the areas of cracks and interstices are unimportant. In fine-grained sediments total surface areas have been measured by gas-adsorption methods (Nelson and Hendricks, 1943; Kulp and Carr, 1952; Diamond and Kinter, 1958). Kulp and Carr's measurements for deep-sea sediments gave areas of between 5 and 40 $m^2/gm$ of dry sediment.

An approximate total surface area for a sediment with particles predominantly of sand and silt size can be computed by assuming all particles to be quartz spheres. Areas are computed for each weight fraction of a grain size analysis, using as a measure of particle size the mean diameter of the fraction. For clay-sized particles it probably is unrealistic to assume a spherical shape, so an area computation for a sediment with an appreciable clay content will be no more than a crude minimum approximation. An additional problem with clayey sediments is that it is difficult and rather time-consuming to carry a grain size analysis beyond the 10 phi diameter limit; hence ordinarily it is not done, yet an appreciable surface area may be present in the fraction finer than 10 phi diameter.

Surface areas for 109 natural sediment samples were computed in this manner. This group of samples included a number with appreciable clay contents. Grain sizes in the finer fractions were determined by the pipette method, and the analyses were stopped at the 10 phi limit. The weight frac-

[1] Manuscript received October 24, 1959.

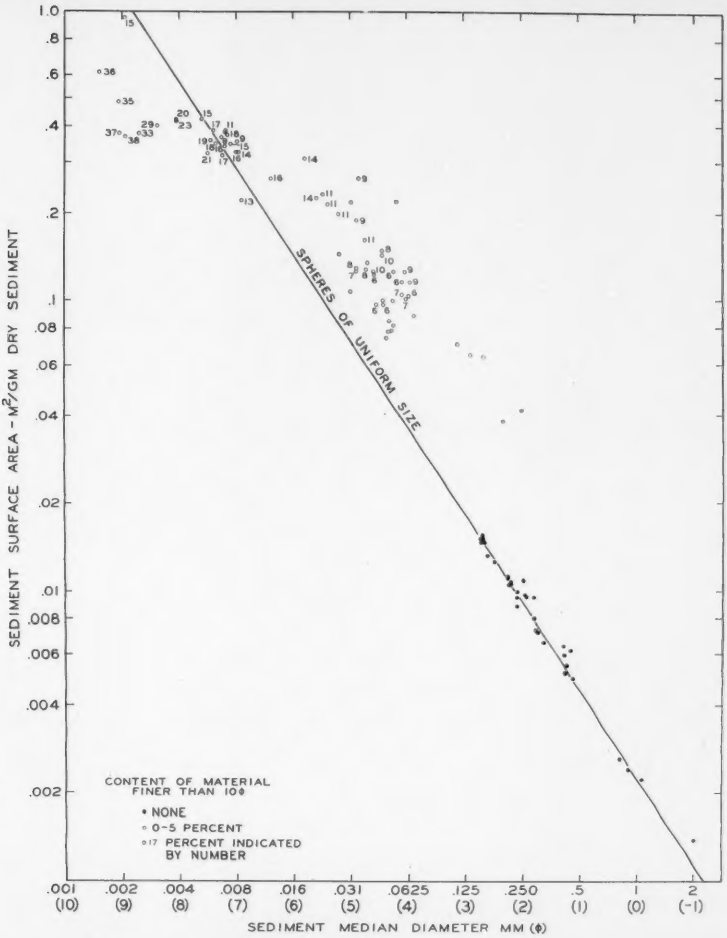


FIG. 1.—Grain surface areas for natural sediments computed from size analysis fractions assuming spherical particles.

tion finer than 10 phi for each sample is indicated on figures 1 and 2, if it amounts to more than 5 percent of the sample.

Sediment surface area data are useful measured both in terms of area per unit mass of dry sediment, and area per unit volume of wet sediment; hence figure 1 shows the data for the former case and figure 2 for the latter case. The sediments used in this study were obtained at or near the sea floor surface, so the area-per-unit-volume data are typical for unconsolidated sediments.

Data for surface area per gram of dry sediment against median diameter can best be considered in relation to the surface area for spheres of uniform size (fig. 1). All sediment data should lie above the curve representing surface area per gram for spheres of

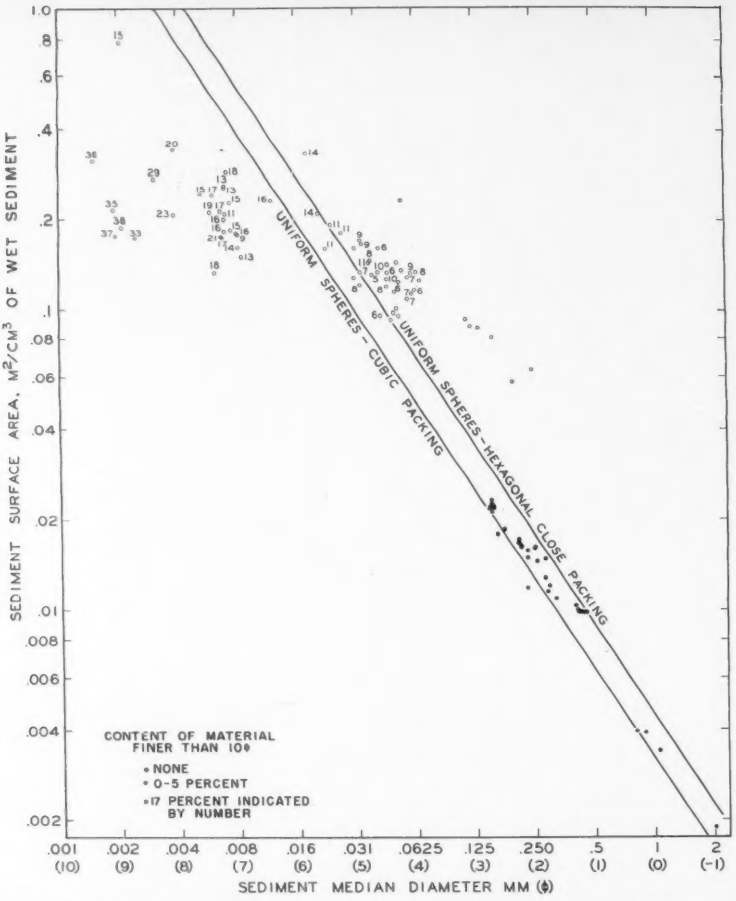


FIG. 2.—Grain surface areas for natural sediments computed from size analysis fractions assuming spherical particles.

uniform size, for it represents the minimum possible area per unit weight. The sand samples (between 0.125 and 2 mm median diameter) lie close to this curve, and a few of these are slightly below the curve probably because of computational errors. The silty sediments lie above the sphere curve because of the presence of appreciable amounts of material finer than the median diameter. The data for fine silts and clays fall near the curve or below it. Obviously these values are much too low.

Data for surface area per unit volume of wet sediment can be considered in relation to curves representing spheres of uniform size in cubic packing and hexagonal close packing (fig. 2). Sands generally lie between these two curves, and a curve midway between these two curves might well be used to predict surface areas for sands. Sediments in the silt median diameter range lie near but generally above the curve for hexagonal close

packing. Clay sediments should have areas falling considerably above both curves, and the fact that the fine silt and clay sediment data do not have such areas points out the considerable surface area contributed by particles finer than 0.001 mm (10 $\phi$).

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 490–495
FIGS. 1–4, SEPTEMBER, 1960

# WOODS HOLE RAPID SEDIMENT ANALYZER[1]

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## ABSTRACT

A recording rapid sediment analyzer has been developed at the Woods Hole Oceanographic Institution from the method and technique described by Appel (1953). The system measures pressure changes induced in a column of water by sediment settling through a measured distance. Mechanical analysis of sediment samples whose sizes are between coarse silts and fine gravels can be made. The authors estimate that one operator could complete curves for a maximum of 150 sand samples in a day.

Reproducibility of results is excellent. In some of the tests there was no variation of the standard deviation around an average particle diameter, at least to the fourth decimal place. A comparison of mechanical analysis by sieving and settling is given, and a method for determining shape is described.

Studies which required size frequency analysis of many hundreds of sand samples stimulated development of a rapid sediment analyzer at Woods Hole Oceanographic Institution. This device has been completed since March 1958, and many size frequency analyses have been made and examined. The writers believe that this procedure has had a reasonable test and wish to present both the technique and an analysis of the method at this time.

Excellent summaries of the development of size frequency analyses by sedimentation methods can be found in reports 4 and 7 prepared by a group of collaborating agencies (U. S. Engineer District, St. Paul, and Hydraulic Laboratory, S.U.I., 1941 and 1943) as well as in various text books (Krumbein and Pettijohn, 1938 for example). Therefore, the writers will not attempt a historical survey of the literature on this subject but leave it to the reader to compare the operation of this device with others of interest to him.

The sediment analyzer developed at Woods Hole (figs. 1 and 2) was designed and built by Mr. Geoffrey G. Whitney, Jr. to exploit a piezometric measuring method developed by Appel (1953). In brief, the pressure of a column of water is measured at two points one meter apart by a bellows within a pressure-tight case.

When a charge of sediment is introduced at the top of the water column, the pressure will increase at the bottom of the column, and as the sediment settles past the pressure sensing hole 1m down the tube, the pressure measured at the bottom hole will return to its original value before sediment was added. If a population of sediment sizes is present, then the rate of change of pressure represents the frequency distribtuion of sediment velocities in the population because the largest grains or the most spherical grains settle fastest and pass out of the measured column first.

## DESCRIPTION AND OPERATION

Two tubes have been developed at Woods Hole. One tube measures sediment fall velocity over a 2 m fall and the other over a 1 m fall. Two holes through each tube either 1 m or 2 m apart transmit the pressure difference by means of glass and rubber tubing to a bellows enclosed in a case (fig. 2). The top hole transmits pressure to the inside of the pressure case (outside of the bellows) and the bottom hole to the inside of the bellows. When pressure at the bottom hole is increased, the bellows is extended. A very sensitive differential transformer mounted on the bottom of the bellows measures the amount of bellows movement, and the signal is sent through an amplifier to a recording device. Either a Sanborn recorder or a Speedomax recorder is used at Woods Hole. It would be possible to have holes at various levels down the tube for shorter fall courses if so desired.

The sediment is introduced into the

[1] Contribution No. 1073 from the Woods Hole Oceanographic Institution. Manuscript received October 26, 1959.

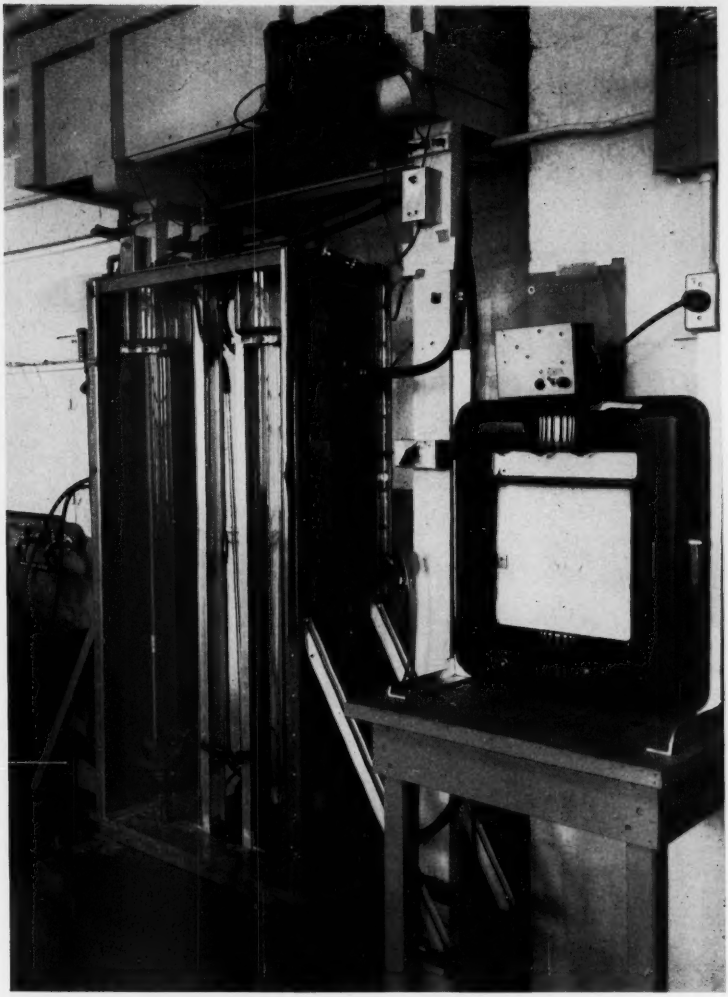

FIG. 1.—Woods Hole Rapid Sediment Analyzer.

column by first placing the charge on a closed knife gate which is submerged at the level of the top hole. After the sediment is wetted, it can be released by opening the gate. The sediment is pushed off or wiped off the gate by a rubber ring as the gates open. The gates are held closed against a spring by a charged solenoid. When the circuit is broken, the gates are opened smoothly and a timing mark put on the record at the same time.

The hot water storage tank permits dissolved gas in the water to be driven off, the room temperature storage tank is simply a

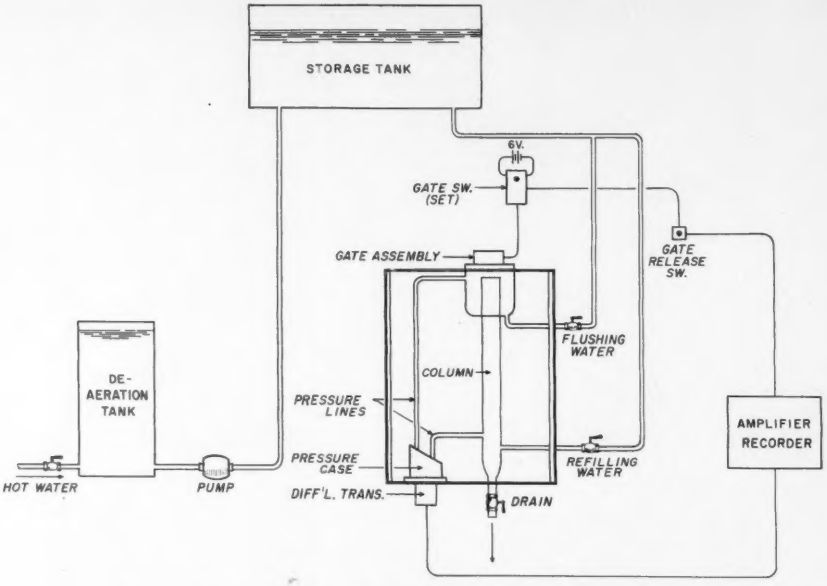


FIG. 2.—Schematic diagram of Woods Hole Rapid Sediment Analyzer.

reservoir needed to fill the tube. This tank is not necessary if one is willing to wait for the hot water to cool in the storage tank.

The writers recommend either lucite or some other plastic for tubing because glass tubes prepared for us, required the skill of an optical company in drilling the holes and making the top edge exactly normal to the axis of the tube. The 2 m tube was developed to measure sediments with a maximum diameter of 2 to 3 mm. Ordinarily, the system is not expected to be useful for analyzing sediment finer than 0.04 mm.

No particular care is required to prepare the samples for analysis. Splits of approximately 5 gm are optimum size, but charges as large as 25 or 30 gm and as small as 2 gm can be used. One does not need to know the sample weight. Small charges of sediment were introduced into the tube in order to find the smallest sample which would cause the marking pen to register. The recorder showed a deflection when 0.05 gm were dropped as the initial charge. If, however, the system was already recording and the

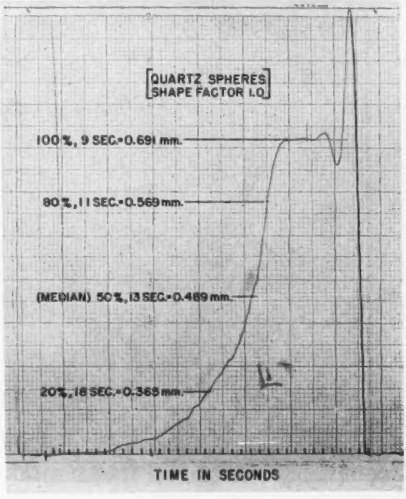


FIG. 3.—Record from Woods Hole Rapid Sediment Analyzer.

TABLE 1.—*Data pertinent for evaluation of the reproducibility of the method*

| Sample No. | Number of times analysis was repeated | 20th percentile | | 50th percentile | | 80th percentile | |
|---|---|---|---|---|---|---|---|
| | | $\bar{x}_{20}$ | $\sigma_{20}$ | $\bar{x}_{50}$ | $\sigma_{50}$ | $\bar{x}_{80}$ | $\sigma_{80}$ |
| 91-1 | 8 | 0.941 | 0.0262 | 0.893 | 0.0169 | 0.81 | 0.0154 |
| E-106 | 9 | 0.837 | 0.0123 | 0.789 | 0.0126 | 0.741 | 0.0088 |
| B-103 | 8 | 0.718 | 0.0093 | 0.667 | 0.0000 | 0.402 | 0.0161 |
| A-26 | 10 | 0.573 | 0.0047 | 0.524 | 0.0109 | 0.300 | 0.0080 |
| A-25 | 15 | 0.570 | 0.0116 | 0.520 | 0.0105 | 0.312 | 0.0115 |
| D-98 | 8 | 0.408 | 0.0047 | 0.345 | 0.0056 | 0.294 | 0.0039 |
| D-97 | 8 | 0.407 | 0.0050 | 0.342 | 0.0000 | | 0.0000 |
| F-107 | 8 | 0.152 | 0.0039 | 0.124 | 0.0025 | | 0.0000 |

Samples arranged in descending order of size.

Where $\bar{x}_{20}$ is the arithmetic mean for the 20th percentile in millimeters, and $\sigma_{20}$ is the standard deviation around the mean for the number of runs made.

original inertia overcome, the recorder reacted to 0.01 gm. Inasmuch as the average charge used is approximately 5 gm, it is believed that the system will detect pressure changes caused by 0.2 percent by weight of 5 gm samples.

The record obtained (fig. 3) can be treated in a variety of ways. Time is measured on the ordinate, and from this one can compute velocity or look up size in tables (Zeigler and Gill, 1959). Percentiles are determined by use of a Gerber variable scale, which can select 100 equal divisions of a line by a single setting.

## REPRODUCIBILITY OF THE METHOD

An effort to discover the reproducibility of the method was made. A single operator completed all phases of the following procedure. Eight different samples having median diameters between 0.893 mm and 0.124 mm were chosen. Each sample was dropped through the settling tube, collected, and dropped again from eight to fifteen times. The samples were weighed after each collection to determine if material was lost. Each run was graphed by the recorder and the various percentiles computed and compared. The results (table 1) indicate that the standard deviation around the same percentiles nowhere exceeds 0.0262, and that the standard deviation is often negligible to the fourth decimal place.

No attempts have been made to determine the maximum speed at which samples can be analyzed by a single operator or by a combination of operations. However, Mrs. Barbara Gill completed 80 curves of beach sand and computed the median diameters and sorting coefficients on 60 of these in a normal working day. The system is certainly faster than any other known to the authors.

This procedure for determining particle size is subject to many uncertainties and limitations. The usefulness of a mechanical analysis depends on what is required of it. The reliability of mechanical analyses by sedimentation methods in general is dependent upon several assumptions and upon either estimation of or tolerance of certain sources of error. Uniform viscosity and uniform density throughout the settling column and lack of convection currents within the column are generally assumed. These assumptions are reasonably met with the Woods Hole Rapid Sediment Analyzer unless the machine is placed where severe temperature gradients form in the settling column.

Sources of error arising from mass properties of the sediment are more difficult to assess. One may apply a correction for wall proximity to the fall velocity of a single particle (Krumbein and Pettijohn, 1938, p. 91; Schulz, Wilde, and Albertson, 1954, p. 40), but it is difficult to apply this correction to a population of particles settling at the same time. In general, the correction is omitted because it is thought to be small. The same philosopy applies to hindered settling, that is, interference between the grains within a settling population. Ordinarily this correction is omitted, but in certain engineering problems such as sewage settling it is imper-

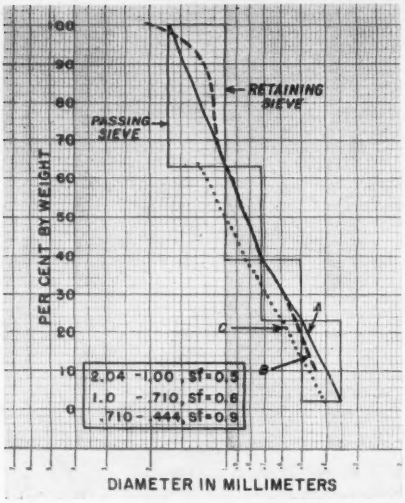


FIG. 4.—Comparison of curve from WHRSA and sieving.

ative to estimate hindered settling (Camp 1946, figs. 3 and 4); also corrections are applied in some clinical measurements (Kermack, McKendrick and Ponder, 1928) and in biological measurements in the centrifuge. Effects of concentration may influence sediment size to a greater degree than is usually supposed, particularly where the sediment is fine and well sorted. This effect, of course, is not unique to the Woods Hole Analyzer but would affect the results of any analysis made by settling. A few preliminary experiments indicate that the effects of concentration should be examined further. Results tabulated on Table 1 indicate that the variation caused by all of the above errors taken together for any one sample is small.

More serious errors arise which are related to other properties of a sediment population, such as sediment samples having a spectrum of specific gravities, that is, being multimineralic. In general, our settling velocities are analyzed as if the sediment samples have the specific gravity of quartz, 2.65. One does not usually attempt to make settling analyses of sediment samples which have noticeably large quantities of heavy minerals, yet this decision depends upon the investigator. One might, for example speculate on the effect of a spectrum of specific gravities on the number of modes.

Most analyses are made as if all fractions of a sediment sample have the same hydrodynamic shape. This assumption, perhaps, leads to more serious difficulties than assuming a monomineralic assemblage, and some readily useful system for determining shape spectra is well worth developing. One way to make an evaluation of the spectrum of shapes present in a sample is to compare the results of a settling analysis with the results obtained by screening the same sample.

One knows that the intermediate axis of sediment caught on a screen is coarser than the retaining screen size but finer than the screen which passed it. Therefore, the sizes of sediment caught on any one screen should fall between the sizes of the retaining and passing screens (fig. 4, line A). If one assumes a uniform distribution of sediment size within any one sieve interval and if one also assumes that the size of the screen opening rather than some function of the diagonal of the opening is a true measure of the size, then curve A on figure 4, connecting the screen sizes, represents the mechanical analysis of that sample determined by screens.

It is possible to fit the cumulative frequency curve obtained by settling for this same sample to the curve obtained by sieving by choosing shape factors to adjust the settling velocity curve to the assumed correct position obtained by screening.

Corey (1949) shape factors of 0.5, 0.6 and 0.9 are needed to fit the settling Curve B to the Curve A of the same sample obtained by screening. The Corey factor[2] $c/\sqrt{ab}$ is used because of the availability of drag coefficients and Reynolds numbers in terms of this factor (Shultz, Wilde, and Albertson, 1954, table 15) and because of the availability of settling velocities prepared in terms of shapes from these drag coefficients and Reynolds numbers (Zeigler and Gill, 1959).

One could likewise fit a curve obtained by settling to other assumptions concerning screening; for example, that any one screen can pass material whose intermediate axis is as much as 1.414 times the screen opening

[2] $a$ = the long axis of a grain, "b" the intermediate axis, and "c" the short axis.

(the diagonal). Perhaps the mid-point of the screen interval is a better estimate of size, Curve C. In this particular case, a Corey factor of about 0.6 will adjust the curves for the entire sample.

At any rate, some assumption of shape is necessary in the procedure for determining sediment size from settling velocities where Reynolds Numbers are less than 0.1. It has been found that Cape Cod sands are very rough and tend to have much lower shape factors than one would get by direct measurement of the axial ratio. Roughness was not included in the computations for particle settling velocity; therefore particles which are rough fall slower than predicted and appear to be less spherical (as defined) than they are. Not only does this point to another source of uncertainty in mechanical analysis determined from settling velocity, but it illustrates that experience with a region helps the investigator choose reasonable values for shape.

In spite of the sources of uncertainty in obtaining sediment sizes from settling velocities, the technique will probably continue to be used. There is a whole class of problems in sediment transport or deposition where the settling velocity itself is a necessary parameter. In problems where sedimentation techniques are valid and desired, the Woods Hole Rapid Sediment Analyzer will be useful.

ACKNOWLEDGMENTS

The authors wish to express their thanks to Mr. Robert Walden of the Electronics Department and Mr. James Gifford of the Machine Shop for their contribution in development of this instrument.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 3, PP. 496–500
SEPTEMBER, 1960

## NOTES

# REMARKS ON THE APPLICATION OF FRIEDMAN'S STAINING METHODS[1]

KARL H. WOLF[2] AND S. ST. J. WARNE[3]

Extensive applications of the tests advocated by G. H. Friedman in the Journal of Sedimentary Petrology V. 29, No. 1, p. 87–97, 1959, for the differential staining of the minerals calcite, aragonite, high-magnesian calcite, dolomite and gypsum etc., in a variety of recent and ancient sediments have yielded some interesting results.

The staining techniques are applicable to single grains, and to unconsolidated carbonate sands, as well as to polished surfaces of well cemented rocks including coal. The non-carbonate minerals even when fine grained remained completely unaffected by the staining solutions, despite their close contact with carbonates, that is there is no spreading of the stain from the carbonate minerals across the material not effected by the staining solutions, except where gross overstaining took place. The minute structures of recent foraminifera were extremely well defined since they consisted of high-magnesian calcite which contrasted sharply with the surrounding pure calcite cement.

Friedman's scheme No. 1, using Alizarin Red S in acid and alkaline conditions and Feigl's solution, proved very reliable and of wide application. It was first applied with complete success to forty single carbonate specimens from the available reference collections, the results being confirmed by chemical analysis or differential thermal analysis. It was then applied to unconsolidated and consolidated carbonate sediments and individual specimens of *Lithothamnion* and foraminifera. Specimens of foraminifera and *Lithothamnion* from Heron Island, Great Barrier Reef and Lord Howe Island proved to be of high-magnesian calcite.[4]

Many of the forams showed secondary filling of their chambers with pure sparry calcite which resulted in a definite color contrast when stained. Two types of *Lithothamnion* colonies, one a quiet water branched type and the other a turbulent water massive encrusting variety, on staining were both found to be dominantly composed of high-magnesian calcite.

Recent shallow water unconsolidated carbonate sands from the same localities proved on staining to consist of grains predominantly of aragonite and high-magnesian calcite with lesser amounts of calcite. Similar results were obtained with beach and dune carbonate sands. Consolidated recent beach limestone from Heron Island proved to be of similar mineralogical composition as their unconsolidated equivalents.

Devonian *Amphipora*, stromatoporoids, reef fragments and pisolitic limestones from the Nisku formation in Alberta, Canada on testing were shown by staining to be of pure calcite. It is noteworthy concerning the pisolitic limestone that relatively small amounts of gypsum present in the calcite matrix, previously overlooked were easily distinguished on staining. The application to subsurface correlation by the use of this rapid and reliable method of identifying gypsum, even when in small amounts, need hardly be stressed further.

Numerous specimens of lithic sandstones and conglomerates from the Cunningham formation (Devonian) of New South Wales, consisting of igneous, shale, limestone and coral fragments, with secondary pyrite, on staining showed admirably the calcite cement as well as the limestone and coral fragments.

Many occurrences of carbonate minerals in veins and nodules in the Permian coals of

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[2] Department of Geology, The University of Sydney, Australia.
[3] Department of Geology, The University of New South Wales, Australia.

[4] Chemical analyses are in progress which are intended to be published at a later date.

New South Wales also were tested with excellent results. The blocks of coal first were highly polished for microscopic examination in reflected light, and then were stained. The preparatory etching and subsequent staining in no way affected the quality of the polish on the coal blocks.

From the above survey we consider that this staining scheme is extremely dependable and of great application, ranging from rapid field work to detailed laboratory examinations on single carbonate occurrences, recent to ancient sediments, and both consolidated and unconsolidated materials.

## COMMENTS ON "SMALL SCALE CROSS-LAMINATION IN LIMESTONES"[1]

KARL H. WOLF
University of Sydney, Sydney, N.S.W.

Harbaugh (1959) states in his paper that ". . . the presence of cross-lamination indicates that the limestone is a detrital deposit and not a chemically or biochemically-formed limestone which has been precipitated in place." Because this was part of his conclusion some remarks may be justified regarding the use of "detrital" and "in place" and regarding his conclusion based on the cross-lamination in general.

The term detrital (and clastic) is generally used in two ways. First, it includes all material that appears to be transported or abraded, and even skeletal constituents, pellets, oolites, etc. (Folk, 1959). Secondly, a detrital or clastic rock is defined as "a rock made up of the debris of other rock" (Rice, 1955).

Considering Harbaugh's statement it seems possible that the calcareous material could have been derived from another rock. On some of the Great Barrier Reef islands for example, semi-consolidated to well-cemented skeletal calcarenites and calcirudites are presently eroded and abraded to form part of the calcilutites and calcarenites in the lagoons and deeper waters. Turbulence in the lagoons cross-laminates the sediments.

On the other hand, physico-chemical and /or biochemical processes can cause the precipitation of lime mud, and with or without the assistance of organic activity this lime mud can subsequently be agglutinated into pellets, oolites, etc. Depending on the environment (turbulence, water depths, etc.) part or all of this material may remain in place or may be transported. Cross-laminated deposits could be accumulated in agitated environments. On some of the Lau Group islands, Fiji, the writer has seen small and large shallow water areas where the calcilutite is shifted back and forth by the tidal waters within the area of formation. Long transportation distances are not necessarily required to form cross-lamination. The question arises: what is the minimum transportation distance for any material, and what is the minimum agitation required, to form cross-lamination? To what extent can one use the term "in place," especially when examining small-scale structures? Strictly speaking, only reef-forming organisms and cement, for instance, will be found in place. Any calcareous material not formed on the surface of deposition itself undergoes at least some transportation, however small the distance may be, before final accumulation.

Therefore, both types of sediments, abraded rock material as well as material formed within the area of deposition by chemical and/or biochemical processes, can result in a cross-laminated limestone with or without small-scale ripple marks.

The possibility of organisms reworking sediments presents another difficulty. Lime mud formed by chemical and/or biochemical processes may after deposition (either in place or after transportation) form a structureless deposit or a laminated or cross-laminated deposit. Gastropods and other mud-eating organisms may rework the lime

[1] Manuscript received June 1, 1960.

mud to form faecal pellets (Wolf, 1960). Depending on the environmental conditions, these pellets remain in place or are transported to another site of deposition.

Under other conditions (Illing, 1954) the lime mud is aggregated into inorganically-formed pellets, oolites, pisolites, and these in turn frequently into lumps.

These are only some examples to illustrate the difficulty in interpreting the genesis of limestones. The presence of cross-lamination alone in limestones cannot be used to interpret the rock as of detrital (if *detrital means detritus of older rocks*) origin, and a possible chemical or biochemical origin cannot be ruled out.

The presence of cross-lamination indicates that some mechanical process was involved during the final stage of deposition. The lack of cross-lamination does not exclude mechanical processes, because lime mud, for instance, can be mechanically transported to a quiet-water environment to form a structureless accumulation.

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## METHOD FOR MOUNTING SILT-SIZE HEAVY MINERALS FOR IDENTIFICATION BY LIQUID IMMERSION[1]

CHARLES W. SPENCER[2]
U. S. Geological Survey, Middletown, Connecticut

In geologic studies involving fine-grained unconsolidated deposits it may be advantageous to determine the mineralogy of silt-size heavy minerals. Recent studies concerned with the diagenesis of terrestrially derived clay minerals upon reaching the marine environment show clearly how difficult it may be to infer the source of clayey marine sediments solely from determination of clay mineralogy. Most "clays" contain an appreciable amount of silt-size, and sometimes coarser, nonclay minerals. The identification of the nonclay minerals, and in particular the heavy minerals, may provide valuable information as to the provenance of clayey sediments. However, it is acknowledged that the clay minerals and nonclay minerals of some "clays" may have had different source areas.

Determination of the mineralogy of nonclay minerals in "clays" formed by the weathering processes may provide exceedingly valuable information on degree of weathering and sourse material(s). The heavy-mineral content may be particularly significant in weathered "clays," owing to considerable variations in stability of different heavy minerals.

The author made a number of heavy-mineral analyses while engaged in research work in the laboratories of the Section of Clay Resources and Clay Mineral Technology of the Illinois State Geological Survey. In the course of this work a method of mounting silt-size heavy-mineral grains on microscope slides with flexible collodion was developed. This technique greatly facilitated optical identification of very fine-grained heavy-mineral suites.

Most sedimentary petrologists are well acquainted with the standard techniques for separation and identification of sand-size heavy minerals. The separation and identification of heavy minerals in the silt-size

[1] Publication authorized by the Chief, Illinois State Geological Survey and the Director, U. S. Geological Survey. Manuscript received October 31, 1959.

[2] Present Address: 951 North Auburn Ave., Farmington, New Mexico.

range, however, presents a variety of problems. Heavy-liquid separations with a separatory funnel are difficult, owing to the relatively slow settling rate of very fine-grained heavy minerals; also, light minerals may tend to buoy up the heavy minerals. Further, the heavy minerals may settle on the sides of separatory funnels. Lightly tapping the side of the funnel will help to shake down those heavy minerals that have been buoyed up by light minerals, and periodic gentle circular stirring with a thin glass rod will facilitate movement of the heavies to the bottom of the funnel. If a centrifuge is available it may be advantageous to centrifuge the samples.

When making heavy-mineral separations of "clays" that have a high percentage of material in the clay-size range it may be desirable to remove the bulk of this clay-size material in order to facilitate heavy-mineral separation of the silt-size fraction. This may be done conveniently by means of the standard method of repeated decantation.

After heavy-mineral separation is practically complete, the heavy separate is placed in a suitable receptacle, such as a watch glass, preferably 3 to 4 inches in diameter. The heavy liquid is removed from the minerals by washing with an appropriate solvent such as acetone or alcohol. The solvent is directed on the grains in a thin stream by means of a wash bottle. The watch glass is then agitated gently with a horizontal-circular motion, which moves the grains to the center of the glass; the solvent and solute may then be carefully poured off and the operation repeated as many times as necessary to remove all traces of the heavy liquid. The separate may then be dried and weighed.

Microscope-slide mounting, and subsequent identification with polarizing microscope, of silt-size minerals presents several problems. Of course the minerals may be mounted in the usual manner by imbedding in suitable transparent cement. However, the cements generally have a relatively fixed refractive index. It was found that frequently the only way to identify or differentiate some of the very-fine silt-size minerals was by differences in their indices of refraction. This is caused primarily by the fact that in very fine-grained minerals intensity of color, such as that of pleochroism and natural mineral color, may decrease with decreasing grain size to the point of being indeterminate; more important, interference figures usually are fairly difficult, and in some instances impossible, to obtain, even when grains are correctly oriented. Therefore, it becomes desirable to have some method by which a given separate may be immersed in more than one refractive-index liquid, if necessary, in order to determine approximate indices of refraction of unknown minerals and to differentiate, during grain counts, those minerals that have very similar characteristics except for differences in index of refraction. For instance, it may be desirable to differentiate between various pyroxene and (or) amphibole minerals. This may be done in the usual manner by placing the grains on a slide and immersing in a liquid of suitable refractive index. However, if it should be necessary to try more than one liquid, a new slide must be prepared for each index liquid used. In some instances there may not be enough of the heavy-mineral separate for many mountings. If the grains could be attached to the slide in such a manner that most of the surfaces of the grains are exposed, then several immersion liquids might be used on one mineral mount. The author became interested in finding some suitable method of attaching silt grains to microscope slides and several mounting media were tried, including kollolith and Canada balsam, but either the grains would not stick properly or would become too deeply embedded. W. A. White of the Illinois State Geological Survey (personal communication, 1955) suggested using flexible collodion. This medium was tried and found very suitable. It can be obtained at most drug stores, and its application to a microscope slide requires no special equipment.

The grain-mounting procedure suggested is as follows: Dip a glass rod in a bottle of U.S.P. flexible collodion and smear a thin, uniform coating of the collodion on a clean microscope slide, allow about 20 seconds for the surface to become tacky, and then gently dust silt-size grains onto the flexible collodion coating. If the flexible collodion is applied too liberally small bubbles may form in the film; if applied too sparingly, it

remains tacky for only a few seconds. When this mounting method is used for the first time it may be desirable to prepare two or three "practice" mounts. Let the flexible collodion dry for about 30 minutes then apply an appropriate immersion liquid and cover glass. Examine heavy minerals under a polarizing microscope. Magnification in the order of 400 to 450 diameters seems to be suitable for identification and grain counts of most silt-size material. If a large percentage of the sample is at the fine end of the silt-size range it may be desirable to go to higher magnifications. To change the immersion liquid, incline the prepared slide over a dish or beaker and direct a thin stream of ethyl alcohol on it by means of a wash bottle. The cover glass will fall off and the immersion liquid will be removed. Quickly shake off any excess alcohol and blow gently on the slide to evaporate the alcohol rapidly before it has time to affect the collodion film. Acetone, though a slightly better solvent for most immersion liquids, reacts fairly rapidly with the collodion, causing the film to fog and pull away from the microscope slide. After the slide is allowed to dry for several minutes, a different index liquid may be used.

PROCEDURE

1.—Remove clay-size fraction, if necessary, by repeated sedimentation and decantation.
2.—Separate heavy-mineral fraction with centrifuge or separatory funnel.
3.—Weigh separates.
4.—Examine under high-power binocular microscope; note distinctive characteristics of mineral suite.
5.—Smear film of U.S.P. flexible collodion on microscope slide.
6.—Allow about 20 seconds for film to become tacky.
7.—Dust mineral grains on tacky film.
8.—Allow approximately 30 minutes to dry.
9.—Immerse grains in suitable index liquid and add cover glass.
10.—Examine under polarizing microscope (generally 400 to 450×).
11.—If desirable to examine grains in another index liquid:
    a.—Wash cover glass and immersion liquid from grain mount with ethyl alcohol, using wash bottle.
    b.—Dry at room temperature as rapidly as possible.
    c.—Repeat step No. 9.

The index of refraction of flexible collodion is slightly variable, but higher than 1.51 and lower than 1.53. Owing to this moderately low index the contrast between the film and high index liquids, such as those higher than 1.70, may appear quite strong; however, this contrast usually causes little difficulty.

The most advantageous liquid to use for grain counts is determined by the particular mineral suite under study, and no specific liquid index can be recommended; however, in studying heavy minerals it is obviously desirable to use a liquid having an index of approximately 1.60 or above, since most heavy minerals have indices higher than this. The author found immersion liquids ranging in index from 1.63 to 1.68 suitable for most heavy-mineral studies.

# NOMINEES FOR S.E.P.M. PRESIDENT, 1961–1962

LAURENCE L. SLOSS

Professor of Geology, Northwestern University, Evanston, Illinois
Born, August 26, 1913, Mountain View, California

*Academic Training*

Stanford University, B.A., '34
University of Chicago, Ph.D., '37

*Experience*

| | |
|---|---|
| 1937–46 | Montana School of Mines (instructor, assistant professor), and Montana Bureau of Mines and Geology (geologist) |
| 1947– | Northwestern University (assistant, associate, professor) |
| 1942–50 | (Summers) Carter Oil Co., Phillips Petr. Co., U.S.G.S. (structure and stratigraphy, Rocky Mountain Area) |

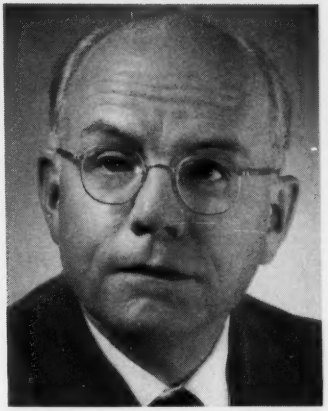
*Publications.*—Stratigraphy and sedimentary petrology

*Professional Affiliations*

American Association of Petroleum Geologists, member
Geological Society of America, fellow
Paleontological Society, member
American Geophysical Union, member
American Association for the Advancement of Science, fellow

*S.E.P.M. Activity* (Member, '48)

| | |
|---|---|
| 1950 | Program Chairman |
| 1954 | Nominating Committee |
| 1955–57 | Research Committee Member |
| 1959 | Co-Chairman Research Committee Symposium |
| 1960–61 | Vice-President |



CURT TEICHERT

Geologist, U. S. Geological Survey, Denver, Colorado
Born, May 8, 1905, Königsberg i. Pr., Germany

*Academic Training*

Albertus University (Königsberg), Ph.D., '28
University of Western Australia, D.Sc., '44

*Experience*

| | |
|---|---|
| 1927–29 | University of Freiburg (assistant) |
| 1930 | U. S. National Museum, N. Y. Museum (Rockefeller fellow) |
| 1931–32 | Danish Expedition to Greenland (geologist) |
| 1933–37 | University of Copenhagen (paleontologist) |
| 1937–45 | University of Western Australia (research lecturer) |
| 1946–47 | Mines Department of Victoria, Australia (asst. chief geologist) |
| 1947–53 | University of Melbourne (senior lecturer) |
| 1953–54 | New Mexico Institute of Technology (professor) |
| 1954– | U. S. G. S. (geologist) |

*Publications.*—Stratigraphy, paleontology (esp. cephalopods), coral reefs

*Professional Affiliations*

Geological Society of America, fellow
American Association of Petroleum Geologists, member
Paleontological Society, member (correspondent, 1951–53)
Paleontological Research Institution
Geochemical Society

*S.E.P.M. Activity* (Associate, '52; Member, '56)

| | |
|---|---|
| 1954–57 | Research Committee (Chairman, 1956–57) |

# NOMINEES FOR S.E.P.M. VICE-PRESIDENT, 1961–1962

KENNETH E. LOHMAN

Geologist, U. S. Geological Survey, Washington, D. C.
Born, September 11, 1897, Los Angeles, California

*Academic Training*

California Institute of Technology, B.S., '29; M.S., '31; Ph.D., '57

*Experience*

1924–26 Certified Laboratory Products (chemist)
1926–31 (Part time) Certified Laboratory Products (chemist)
1931– U. S. Geological Survey (geologist)

*Publications.*—Diatoms, stratigraphy

*Professional Affiliations*

American Association of Petroleum Geologists, member
Geological Society of America, fellow
American Association for the Advancement of Science, fellow
American Geophysical Union, member

*S.E.P.M. Activity* (Member, '60)

1960 Chairman, Technical Services Committee

FRANKLYN B. VAN HOUTEN

Professor of Geology, Princeton University, Princeton, New Jersey
Born, July 14, 1914, New York City

*Academic Training*

Rutgers University, B.S., '36
Princeton University, Ph.D., '41

*Experience*

1939–42 Williams College (teaching and research)
1946– Princeton University (professor)

*Publications.*—Sedimentation and vertebrate paleontology

*Professional Affiliations*

American Association of Petroleum Geologists, member
Geological Society of America, fellow
Sigma Xi

*S.E.P.M. Activity* (Member, '49)

1954–57 Research Committee

## NOMINEES FOR S.E.P.M. SECRETARY-TREASURER, 1961–1962

LAWRENCE C. BONHAM

Research Geologist, California Research Corporation, La Habra, California
Born, June 9, 1920, Springfield, Missouri

*Academic Training*

Drury College, B.S., '42
Washington University, M.S., '48; Ph.D., '50

*Experience*

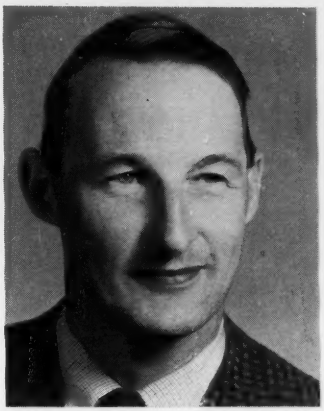
| | |
|---|---|
| 1942–44 | U. S. Coast and Geodetic Survey (photogrammetric engineer) |
| 1944–46 | U. S. Army Air Force (photogrammetrist) |
| 1946 | Missouri Geological Survey (geologist) |
| 1948, 49 | (Summers) Jersey Production Research Co. (geological research) |
| 1950– | California Research Corporation (geological research) |

*Publications.*—Organic geochemistry, structural petrology

*Professional Affiliations*

American Association of Petroleum Geologists, member

*S.E.P.M. Activity* (Member, '59)



LLOYD C. PRAY

Research Geologist, The Ohio Oil Company, Littleton, Colorado
Born, June 25, 1919, Chicago, Illinois

*Academic Training*

Carleton College, B.A., '41
California Institute of Technology, M.S., '43; Ph.D., '52

*Experience*

| | |
|---|---|
| 1942 | Magnolia Petroleum Co. (subsurface geology) |
| 1943–44 | U. S. Geological Survey (pegmatite investigations) |
| 1944–46 | U. S. Navy (hydrographic surveys) |
| 1946–56 | U. S. Geological Survey (rare earth deposits)—Part time |
| 1950–56 | California Institute of Technology (instructor, assistant professor) |
| 1956– | The Ohio Oil Co. (geological research) |

*Publications.*—Stratigraphy, structural geology, petrology

*Professional Affiliations*

American Association of Petroleum Geologists, member
American Institute of Mining, Metallurgical and Petroleum Engineers, member
Geological Society of America, fellow

*S.E.P.M. Activity* (Member, '55)

1959 Leader, Field Conference, Permian Basin Section, S.E.P.M.

# NOMINEES FOR S.E.P.M. COUNCILOR (PALEONTOLOGY), 1961–62

LYMAN D. TOULMIN

Professor of Geology, Florida State University, Tallahassee, Florida
Born, July 4, 1904, Mobile, Alabama

*Academic Training*

University of Alabama, B.A., '26; M.S., '34
Princeton University, Ph.D., '40

*Experience*

| | |
|---|---|
| 1938–42 | Texas A. & M. College (instructor, assistant professor) |
| 1942–45 | Geological Survey of Alabama (geologist) |
| 1945–48 | Birmingham Southern College (associate professor) |
| 1953–54 | Louisiana State University (visiting professor) |
| 1948– | Florida State University (associate professor, professor) |

*Publications.*—Paleontology and stratigraphy

*Professional Affiliations*

American Association of Petroleum Geologists, member
Paleontological Society, member
Geological Society of America, fellow
American Association for the Advancement of Science, fellow
Sigma Xi

*S.E.P.M. Activity* (Member, '54)

GARNER L. WILDE

Paleontologist, Humble Oil & Refining Company, Midland, Texas
Born, September 29, 1926, Spring Creek, Texas

*Academic Training*

Texas Christian University, B.A., '50; M.A., '52

*Experience*

| | |
|---|---|
| 1950–52 | Texaco, Inc. (paleontological assistant) |
| 1952– | Humble Oil & Refining Co. (paleontologist) |

*Publications.*—Paleontological studies on fusulinids

*Professional Affiliations*

American Association of Petroleum Geologists, member
Geological Society of America, member

*S.E.P.M. Activity* (Member, '53)

| | |
|---|---|
| 1955 | Chairman, Steering Committee, to organize Permian Basin Section, S.E.P.M. |
| 1960–61 | President, Permian Basin Section (Non-voting member of the Council) |

## NOMINEES FOR S.E.P.M. COUNCILOR (MINERALOGY), 1961–1962

WILLIAM C. GUSSOW

Staff Geologist, Union Oil Company of California, Calgary, Alberta, Canada
Born, April 25, 1908, London, England

*Academic Training*

Queen's University (Kingston, Ontario), B.S., '33; M.S., '35
Massachusetts Institute of Technology, Ph.D., '38

*Experience*

| | |
|---|---|
| 1945–52 | Shell Oil Company (chief geologist, exploration manager) |
| 1953–55 | Consulting geologist and engineer |
| 1956– | Union Oil Company of California (staff geologist) |

*Publications.*—Differential entrapment and migration of oil and gas, stratigraphy, structural geology

*Professional Affiliations*

Royal Society of Canada, fellow
Geological Society of America, fellow
Geological Association of Canada, fellow
American Association of Petroleum Geologists, member
Society of Economic Geologists, member
American Geophysical Union, member
Canadian Institute Mining and Metallurgy, member

*S.E.P.M. Activity* (Member, '56)

| | |
|---|---|
| 1960 | Co-Chairman Research Committee Symposium |

ALBERT L. KIDWELL

Research Geologist, Jersey Production Research Company, Tulsa, Oklahoma
Born, January 1, 1919, Auxvasse, Missouri

*Academic Training*

Missouri School of Mines, B.S., '40
Washington University, M.S., '42
University of Chicago, Ph.D., '49

*Experience*

| | |
|---|---|
| 1942–44 | U. S. Coast and Geodetic Survey (construction of maps) |
| 1944–47 | Missouri Geological Survey (geologic mapping and economic geology) |
| 1950– | Jersey Production Research Company (research) |

*Publications.*—Recent sediments and occurrence of oil

*Professional Affiliations*

American Association of Petroleum Geologists, member
Geological Society of America, member
Mineralogical Society of America, member
Sigma Xi

*S.E.P.M. Activity* (Member, '59)

| | |
|---|---|
| 1960 | Organized program on Mineralogy |

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## REVIEW

*Historical Geology*, Second Edition, by Carl O. Dunbar, 1959. Pp. xi+500, 406 figs., 11 charts; 7½×10 in., cloth. John Wiley & Sons, Inc., New York. Price $7.95.

The second edition of this historical geology text should be even more popular than its first edition which sold more than 74,000 copies in the past 11 years because, whereas general philosophy and style of the previous edition has been retained, many improvements and additions have been made in this edition.

The most obvious change is the size of the second edition. The page size has been increased to 7½×10 inches which allows for larger and clearer illustrations. The number of words in the new edition is practically the same as in the former edition, but the larger page size has permitted the inclusion of 56 more illustrations. Approximately one-fourth of the illustrations in this edition are new. Many of the illustrations retained from the first edition are more useful now that they are larger.

In the decade since the publication of the first edition, there have been many advances in such fields as geochronology, astronomy, geochemistry, paleontology, and stratigraphy which have added knowledge to the general field of historical geology. Some of these advances are reflected in the more significant changes and additions in the second edition.

The revision of the chapter on the cosmic history of the earth is one of the major improvements which is to be found.

The controversial cloud cover has been removed from all paleogeographic maps which have been redrawn, are larger, and thus are more useful in this edition. It is a disappointment, however, that there is no mention of the idea that island arcs and not bordering land masses may have served as the sources of sedimentation in the Appalachian, Ouachita, and Cordilleran geosynclines as has been pointed out in other recent textbooks.

The stratigraphic correlation charts of the periods of geologic time have been removed from the appendix and placed in the text. Unfortunately, a minor error has resulted in the switching of the Cretaceous and Cenozoic charts, but this and other errors can be easily remedied in subsequent printings.

The new edition also includes material on the origin of life on earth and some consideration of what the earth was like before life began. Advances in human paleontology and other phases of vertebrate paleontology are also included. There is more material on radioactive dating, one result of which is the abandoning of the idea of splitting the Precambrian into two distinct eras, the Archeozoic and Proterozoic.

The familiar plates of invertebrate fossils are retained and are more lucid than ever in their larger size. In addition, three new Cenozoic fauna plates have been added.

Throughout the text more up-to-date figures on such things as mineral production will be noticed. Perhaps the most significant change is the revision of the references at the end of each chapter which will guide the interested student to some of the latest general references on many of the topics which cannot be adequately discussed in a general textbook on historical geology.

While not every teacher of historical geology will find Dunbar's second edition to be the final word in historical geology textbooks, this new edition is a significant contribution to the every-growing field of Historical geology textbooks, and there is every reason to believe that Dunbar's *Historical Geology* will remain among the most popular texts in this field.

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